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Geology of the Tuscarora geothermal prospect, Elko County, Nevada

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ABSTRACT

The Tuscarora geothermal prospect is located at the north end of Independence Valley in northern Nevada. Thermal springs issue from Oligocene tuffaceous sediments near the center of an area of high thermal gradient. The springs are associated with a large siliceous sinter mound and are currently depositing silica and calcium carbonate. Measured fluid temperatures range up to 95 °C, and chemical geothermometers indicate a reservoir temperature of 216 °C. The Independence Valley contains 35- to 39-m.y.-old tuffs and tuffaceous sediments which overlie Paleozoic clastic and volcanic rocks and are overlain by Miocene lava and pyroclastic flows. The rocks have been deformed by normal faults trending north-south and northwest and by folds trending north-south which have been active in the Pleistocene.

INTRODUCTION

The Tuscarora geothermal prospect is located at the north end of Independence Valley, ~80 km north-northwest of Elko, Nevada. Independence Valley is a structural basin between the Independence Mountains on the east and the Tuscarora and Bull Run Mountains to the west and northwest (Fig. 1).

Surface expression of the geothermal resource consists of thermal springs and associated siliceous sinter and travertine deposits. The numerous hot springs extend for 1.4 km along Hot Creek. Deposits include an inactive opaline sinter mound 35 m high and about 1 km long at the southern springs, and a small area of travertine at the northern springs. The northern springs, some of which are boiling, have the greater discharge rate. The springs are depositing both siliceous and calcareous sinter.

The geothermal prospect was leased by AMAX Exploration, Inc., Geothermal Branch in 1977. Their geothermometry indicated a reservoir temperature of 216 °C (Pilkington and others, 1980). The exploration program has included drilling of 38 thermal gradient holes, gravity, aeromagnetic, and electrical surveys (Berkman, 1981), and a deep (1,663 m) test hole.

GEOLOGIC SETTING

Independence Valley is a north-south–trending graben in the northern Basin and Range province. The northern Independence Mountains to the east of the valley consist of allochthonous Ordovician quartzites, shales, and cherts of the Valmy Group (Churkin and Kay, 1967) that are thrust over Paleozoic carbonate rocks (Kerr, 1962; Churkin and Kay, 1967). The allochthonous rocks were eroded and overlain by Mississippian to Permian conglomerate, shale, chert, and quartzite tentatively correlated with the overlap assemblage by Miller and others (1981). The Schoonover Formation (Fagan, 1962) was thrust over or faulted against these rocks in the northern part of the Independence Mountains (Miller and others, 1981).

The Tuscarora Mountains to the southwest of the study area (Fig. 1) consist of Tertiary volcanic and sedimentary rocks overlying Ordovician rocks (Hope and Coats, 1976). To the north, lower Paleozoic limestone and quartzite, Cretaceous intrusive rocks, and a Tertiary porphyritic andesite intrusion are exposed in the Bull Run Mountains (Decker, 1962).

Regional Tertiary volcanism in northwestern Elko County began in late Eocene with eruption of silicic pyroclastics (Stewart and Carlson, 1976). Thick tuffaceous sediments derived from these rocks, and contemporaneous eruptions were deposited during the early Oligocene. Extensional tectonism during the late Tertiary produced Basin and Range structures while volcanic activity continued in the form of lava flows and welded ash-flow tuffs.

STRATIGRAPHY

Paleozoic Rocks

Paleozoic sedimentary rocks crop out to the east and north of Jack Creek (Fig. 2)[1], and underlie the Independence Mountains. The present geologic mapping was extended onto the Paleozoic rocks on a reconnaissance basis in order to determine the distribution of structures, especially high-angle faults, and lithologies that may play a role in controlling the movement of geothermal fluids beneath the Tertiary and Quaternary rocks to the west. A detailed study of the stratigraphy and petrography of the Paleozoic section is beyond the scope of this study; the interested reader is referred to Churkin and Kay (1967), Fagan (1962), and Miller and others (1981) for more detailed discussions.

A massive white to hematite-stained orthoquartzite is exposed on the west flank of the Independence Mountains. The massive quartzite fits the description of the McAfee Quartzite member of the Ordovician Valmy Group (Churkin and Kay, 1967). Dark gray to black cherts and quartzites with thin shalely partings exposed along Schoonover Creek, Boyd Creek, and Jack Creek are also thought to be correlative with rocks of the Valmy Group based on fossils collected by Churkin and Kay (1967). The stratigraphic rela-

[1]Figure 2 is presented on a foldout in this issue.

Geological Society of America Bulletin, v. 93, p. 1264–1272, 6 figs., 1 table, December 1982.

tectonics of the Laramide orogeny, *in* Matthews, V., III, ed., Laramide folding associated with basement block faulting in the western United States: Geological Society of America Memoir 151, p. 355–366.

Dobrin, M. B., 1976, Introduction to geophysical prospecting (3rd edition): New York, McGraw-Hill, 630 p.

Droullard, E. K., 1963, Tectonics of the southeast flank of the Hartville uplift, Wyoming: Rocky Mountain Association of Geologists Symposium on the northern Denver Basin, p. 176–178.

Graff, P. J., Sears, J. W., and Holden, G., 1980, Investigations of uranium potential of Precambrian metasedimentary rocks, central Laramie Range: Department of Energy Report GJBX-22-81.

Hamilton, W., 1981, Plate-tectonic mechanism of Laramide deformation: University of Wyoming Contributions to Geology, v. 19, p. 87–92.

Hills, F. A. and Armstrong, R. L., 1974, Geochronology of Precambrian rocks in the Laramie Range and implications for the tectonic framework of Precambrian southern Wyoming: Precambrian Research, v. 1, p. 213–225.

Hills, F. A., Gast, P. W., Houston, R. S., and Swainbank, I., 1968, Precambrian geochronology of the Medicine Bow Mountains of southeastern Wyoming: Geological Society of America Bulletin, v. 79, p. 1757–1784.

Hills, F. A., Houston, R. S., and Subbarayudu, G. V., 1975, Possible Proterozoic plate boundary in southern Wyoming: Geological Society of America Abstracts with Programs, v. 7, p. 614.

Hodge, D. S., Owen, L. B., and Smithson, S. B., 1973, Gravity interpretation of the Laramie anorthosite complex, Wyoming: Geological Society of America Bulletin, v. 84, p. 1451–1464.

Houston, R. S., 1971, Regional tectonics of the Precambrian rocks of the Wyoming province and its relationship to Laramide structure: Wyoming Geological Association, 23rd Annual Field Conference, Guidebook, p. 19–28.

Houston, R. S. and McCallum, M. E., 1961, Mullen Creek–Nash Fork shear zone, Medicine Bow Mountains, southeastern Wyoming [abs.]: Geological Society of America Special Paper 68, p. 91.

Houston, R. S., and others, 1968, A regional study of rocks of Precambrian age in that part of the Medicine Bow Mountains lying in southeastern Wyoming—With a chapter on the relationship between Precambrian and structure: Geological Survey of Wyoming Memoir 1, p. 617.

Houston, R. S. Karlstrom, K. E., Hills, F. A., and Smithson, S. B., 1979, The Cheyenne Belt: The major Precambrian crustal boundary in the western United States [abs.]: Geological Society of America Abstracts with Programs, v. 11, p. 446.

Jackson, W. H., Stewart, W., and Pakister, L. C., 1963, Crustal structure in eastern Colorado from seismic-refraction measurements: Journal of Geophysical Research, v. 68, p. 5767–5776.

Karlstrom, K., Houston, R. S., and others, 1981, A summary of the geology and uranium potential of Precambrian conglomerates in southeastern Wyoming: U.S. Department of Energy Report DJBK-139-81.

McCallum, M. E., 1974, Dedolomitized marble lenses in shear zone tectonites, Medicine Bow Mountains, Wyoming: Journal of Geology, v. 82, p. 473–487.

McCallum, M. E., Eggler, D. H., and Burns, L. K., 1975, Kimberlitic diatremes in northern Colorado and southern Wyoming: Physics and Chemistry of the Earth, v. 9, p. 149–161.

Newhouse, W. H., and Hagner, A. F., 1957, Geologic map of anorthosite areas, southern part of Laramie Range, Wyoming: U.S. Geological Survey Miscellaneous Field Studies Map MF-119.

Peterman, Z. E., 1981, Dating of Archean basement in northeastern Wyoming and southern Montana: Geological Society of America Bulletin, Part I, v. 92, p. 139–146.

Peterman, Z. E., and Hedge, C. E., 1968, Chronology of Precambrian events in the Front Range, Colorado: Canadian Journal of Earth Sciences, v. 5, p. 749–756.

Press, F., 1966, Seismic velocities, *in* Clark, S. P., Jr., ed., Handbook of physical constants: Geological Society of America Memoir 97, p. 195–221.

Prodehl, C., 1979, Crustal structure of the western United States: U.S. Geological Survey Professional Paper 1034, 74 p.

Prodehl, C., and Pakiser, L. C., 1980, Crustal structure of the southern Rocky Mountains from seismic measurements: Geologic Society of America Bulletin, Part I, v. 91, p. 147–155.

Sales, J. K., 1968, Crustal mechanics of Cordilleran foreland deformation: A regional and scale model approach: American Association of Petroleum Geologists Bulletin, v. 52, p. 2016–2044.

Schilt, S., Oliver, J., Brown, L., Kaufman, S., Albaugh, D., Brewer, J., Cook, F., Jensen, L., Krumhansl, P., Long, G., and Steiner, D., 1979, The heterogeneity of the continental crust: Results from deep crustal seismic reflection profiling using the VIBROSEIS technique: Reviews of Geophysics and Space Physics, v. 17, p. 354–368.

Smithson, S. B., Shive, P. N., and Brown, S. K., 1977, Seismic reflections from Precambrian crust: Earth and Planetary Science Letters, v. 37, p. 333–338.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J., and Hurich, C., 1978, Nature of the Wind River thrust, Wyoming, from COCORP deep reflection data and from gravity data: Geology, v. 6, p. 648–652.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J. E., and Hurich, C. A., 1979, Structure of the Laramide Wind River uplift, Wyoming, from COCORP deep reflection data and from gravity data: Journal of Geophysical Research, v. 84, p. 5955–5972.

Tweto, O., 1975, Laramide (Late Cretaceous–early Tertiary) orogeny in the southern Rocky Mountains, *in* Curtis, B. M., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 1–44.

Tweto, O., and Sims, P. K., 1963, Precambrian ancestry of the Colorado mineral belt: Geological Society of America Bulletin, v. 74, p. 991–1014.

Warner, L. A., 1978, *The Colorado Lineament*: A middle Precambrian wrench fault system: Geological Society of America Bulletin, v. 89, p. 161–171.

Warren, D. H., and Healy, J. H., 1973, Structure of the crust in the conterminous United States: Tectonophysics, v. 20, p. 203–213.

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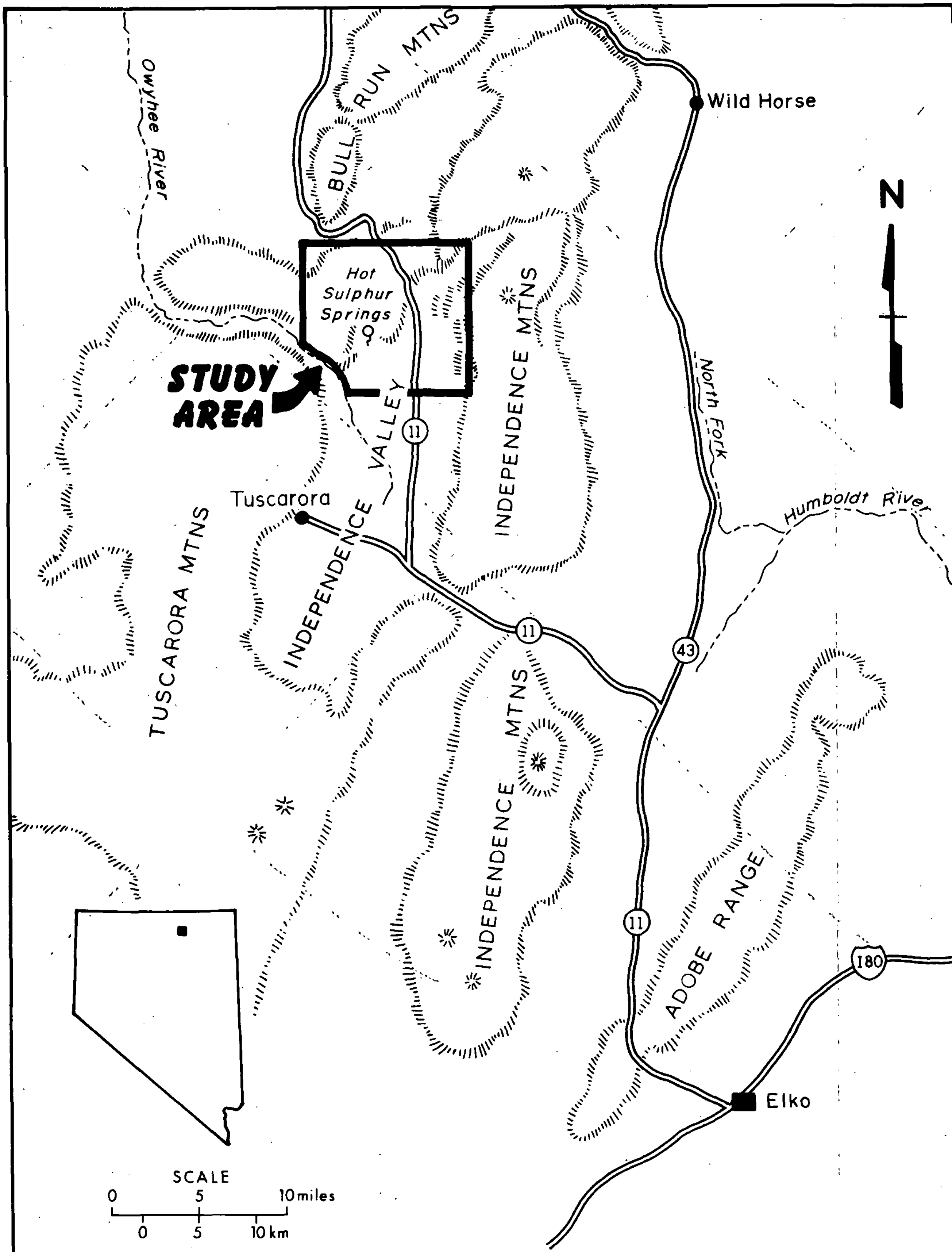

Figure 1. Location map and physiographic setting of the Tuscarora area, Elko County, Nevada.

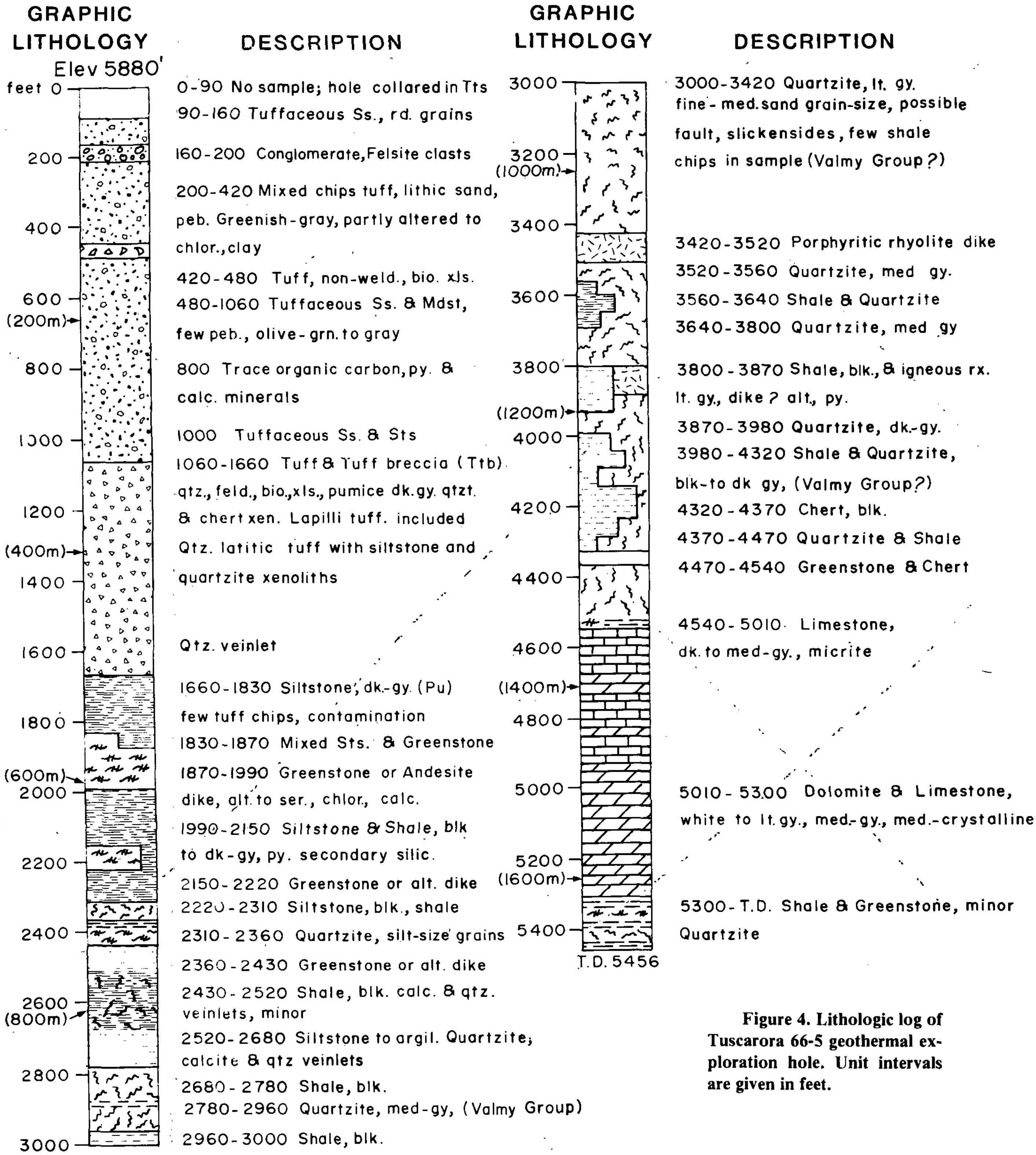

Figure 4. Lithologic log of Tuscarora 66-5 geothermal exploration hole. Unit intervals are given in feet.

tionship between the chert unit and the massive quartzite to the east is uncertain because of the extensive surficial deposits and high-angle faulting between them.

The chert unit is unconformably overlain by conglomerate and argillites of the Mississippian to Permian overlap assemblage (Miller and others, 1981; Roberts and others, 1958). Although the unconformity is well exposed along Schoonover Creek, its location is uncertain to the south and east due to poor exposure and faulting. The correlation of poorly exposed argillite, quartzite, chert, and greenstone between Jack Creek and Marsh Creek is therefore uncertain, and these rocks are here grouped as Paleozoic rocks undifferentiated. Previous investigators have mapped the rocks

TABLE 1. POTASSIUM-ARGON AGE DETERMINATIONS FOR ROCKS IN THE TUSCARORA GEOTHERMAL AREA, ELKO COUNTY, NEVADA

Sample no.	Unit	Material dated	Location (Lat. N, Long. W)	K (%)	Moles/g §Ar^{40}_{Rad} (×10^{11})	§Ar (%)	Age (m.y. ± 1σ)
NVT-32*	andesite (Tal)	plagioclase	41°27′7″ 116°12′35″	0.623	4.458	80	40.8 ± 1.3
NVT-41*	dacitic ash-flow tuff (Ttb)	biotite	41°30′10″ 116°11′26″	7.01	46.924	58	38.2 ± 1.4
NVT-42*	andesite (Tvi)	plagioclase	41°31′7″ 116°10′54″	0.90	2.617	23	16.7 ± 1.1
NVT-43*	welded ash-flow tuff (Tvt)	sanidine	41°31′21″ 116°10′32″	6.30	16.254	52	14.8 ± 0.6
NVT-45*	latitic ash-flow tuff (Ttb)	biotite	41°27′12″ 116°12′0″	6.68	45.440	77	38.8 ± 1.3
Ta 3 †	andesite (Taf)	whole rock	41°30′6″ 116°6′13″	3.87	9.163	56.1	13.6 ± 0.7

* S. H. Evans, Jr., analyst.
† Analysis by Teledyne Isotopes, 1980.
§ Radiogenic argon.
Constants: $\lambda_\epsilon = 0.581 \times 10^{-10}$/yr, $\lambda_\beta = 4.962 \times 10^{-10}$/yr; atomic abundance: K^{40}/K M 1.167×10^{-4}.

between Marsh Creek and Jack Creek as Schoonover Formation (Churkin and Kay, 1967) or overlap assemblage and Schoonover Formation (Miller and others, 1981). These rocks are structurally complex, the lithologies occurring in discontinuous lenses with thick massive quartzite. The Schoonover Formation, however, as mapped by Fagan (1962) north of Jack Creek consists of relatively uniform and laterally continuous well-bedded units with a north-to-northeast dip (Fig. 2). In outcrop and on aerial photographs, the Schoonover Formation appears little deformed, whereas the rocks south of Jack Creek are highly deformed. The base of the Schoonover Formation as originally defined by Fagan (1962) is a fault here interpreted to be the thrust or tear fault exposed along Jack Creek (Fig. 2). The present investigation therefore concluded that Schoonover Formation rocks are not present south of the major fault along Jack Creek.

The Mississippian Schoonover Formation (Fagan, 1962) crops out north of Jack Creek (Fig. 2) and consists of deep-water eugeosynclinal deposits of shale, chert, greenstone, turbidites, and well-bedded quartzite. The base of the Schoonover Formation is a fault exposed along Jack Creek which was mapped as a thrust by Miller and others (1981).

Tertiary Rocks

Vent facies pyroclastic deposits consisting of rhyolitic ash, lapilli, and blocks form a nonstratified, heterogeneous tuff-breccia two miles west of Hot Sulphur Springs (Fig. 2). The deposit includes ash-flow tuffs of limited extent which grade into tuff-breccia. Quartz and biotite phenocrysts are present in the ash-flows. Lapilli to block-sized xenoliths of shale, chert, and quartzite are abundant near the periphery and/or base of the deposit.

The tuff-breccia occurs in a fault-bounded horst, and depositional contacts with other units are rare. In a few exposures, the tuff-breccia seems to overlie or intertongue with the tuffaceous sedimentary unit (Tts). Along Skull Creek, on the west side of the study area (Fig. 2), a small outcrop of Paleozoic chert and quartzite is exposed under the tuff-breccia. If this outcrop is in place, and not a large xenolith brought up by the vent, it would indicate that the tuff-breccia lies directly on the Paleozoic basement at this locality.

A thickness of more than 280 m of tuff-breccia is exposed along Skull Creek within the vent. The geothermal test well, Tuscarora 66-5, penetrated 183 m of a tuff-breccia which is interpreted as the same unit (Figs. 3 and 4)[2]. The tuff-breccia in Tuscarora 66-5 overlies Paleozoic rocks and is overlain by 323 m of tuffaceous sediments (Fig. 4). A similar stratigraphy was encountered 1.5 km to the southeast in Tuscarora 51-9 (Fig. 3; John Deymonaz, AMAX, 1980, oral commun.).

Potassium-argon dates of 38.2 ± 1.4 m.y. and 38.8 ± 1.3 m.y. on biotite were determined for two samples of ash-flow tuff within the tuff-breccia unit (Table 1). This places the tuff-breccia in the age range of the ash-flow tuffs southwest of the study area in the Tuscarora Mountains (Stewart and Carlson, 1976).

A thick, well-bedded unit of tuffaceous sedimentary rocks, volcaniclastic conglomerate, and non-welded tuffs is exposed from northern Independence Valley to north of Deep Creek. A thickness of 323 m of tuffaceous sedimentary rocks was penetrated in the geothermal test well Tuscarora 66-5 (Fig. 4). The lower part of the sedimentary rocks appear to be time equivalent to the tuff-breccia because of intertonguing exposed in outcrop. The pyroclastics (Ttb) were probably a local source of volcanic detritus when the vent was active. Volcanic vents outside of the study area, such as in the Tuscarora Mountains to the south, also contributed to the tuffaceous sediments.

The tuffaceous sedimentary rocks correlate with the Ts1 and Ts2 units of Hope and Coats (1976). A potassium-argon date of 35.2 ± 1.0 m.y. on biotite from a rhyolite ash within the unit was obtained on a sample collected in section 11, along Deep Creek, southwest of Lime Mountain (Schilling, 1965). The sample site appears to be outside the horst and may be part of the upper and younger part of the tuffaceous sedimentary rocks.

[2]Figure 3 appears on the same foldout as Figure 2 in this issue.

Andesite and basaltic andesite lava flows overlie the tuffaceous sedimentary rocks and the tuff-breccia along the Owyhee River and are interbedded with the sediments on the west side of Hot Creek (Fig. 2). The flows are porphyritic, and some are flow breccia (aa flows) and locally columnar jointed where massive. Although the andesite flows clearly overlie tuff-breccia and tuffaceous sedimentary rocks in some exposures, other contact relationships are unclear. The outcrop pattern on the west side of Hot Creek suggests that some andesite flows may be interbedded with the tuffaceous sediments (Fig. 2). Sample NVT-32, an andesite collected in section 14 along the South Fork Owyhee River, has a K-Ar date of 40.8 ± 1.3 m.y. (Table 1). This flow overlies tuffaceous sedimentary rocks to the north; and flows of the same unit, but possibly a little younger, overlie the tuff-breccia unit about 1 km to the northeast. The sample NVT-45 of a welded-ash flow within the tuff-breccia in the same area gave a K-Ar date of 38.8 ± 1.3 m.y. (Table 1). The andesite flows and the tuff-breccia are therefore contemporaneous within the confidence interval of the two dates. Exposures indicate that the andesite flows are overlying the flanks of the tuff-breccia vent area. The andesite flows unit (Tal) of this report correlates with Ta_2 of Hope and Coats (1976).

Basaltic andesite plugs and dikes have intruded the tuffaceous sedimentary rocks along Hot Creek and the tuff-breccia along the vent margin. The age of these intrusions is uncertain, but they are mineralogically very similar to some of the basaltic andesite lava flows.

A unit of porphyritic lava flows consisting of quartz latite and dacite flows overlies the tuffaceous sedimentary rocks in most of the study area (Tvi, Fig. 2). Outcrops are reddish-brown and closely joined where the unit is devitrified. Dark green to black vitric zones are present and, where exposed, flow tops are vesicular. A porphyritic andesite flow containing 7% andesine phenocrysts and a trace of pyroxene is present at the top of the flows. A K-Ar date of 16.7 ± 1.1 m.y. on plagioclase was determined for sample NVT-42 from this flow (Table 1). There is a slight discordance between the bedding in the sedimentary rocks and the base of the lava flows which suggests that this contact is an angular unconformity. The thickness of the lava flow unit varies from about 30 to more than 120 m. Hope and Coats (1976) correlated these flows with the Jarbridge Rhyolite.

Scattered erosional remnants of a poorly welded vitric-crystal ash-flow tuff (Tvt, Fig. 2) overlie the latite and dacite lava flows. The tuff is 60 m or less thick and was extensively eroded before emplacement of the porphyritic andesite flow (Taf, Fig. 2) which overlies both tuff and latite-dacite composition lava flows west of Harrington Creek. A K-Ar date of 14.8 ± 0.6 m.y. on sanidine was determined for sample NVT-43 of the ash-flow tuff (Table 1). A similar K-Ar date of 15.3 ± 0.5 m.y. on sanidine was determined for a welded tuff (McKee and others, 1976) a few miles to the west (lat. 41°33′N, long. 116°17′30″W).

The porphyritic andesite flow along Harrington Creek (Taf, Fig. 2) may be the youngest lava flow in the study area and has been dated at 13.6 ± 0.7 m.y. by K-Ar methods (H. D. Pilkington, AMAX, 1980, written commun.). The andesite flow rests on the dacite-latite lava flows along most of the exposure, but the sanidine ash-flow tuff (Tvt) underlies the andesite where the tuff is preserved.

Quaternary Deposits

Extensive alluvial gravel deposits containing mostly Paleozoic quartzite boulders overlap the Tertiary and Paleozoic rocks around the north end of Independence Valley. These gravel deposits have been subdivided into three units (QTg, Qg, and Qoa, Fig. 2) based on their relationship to the present drainage system, general composition, outcrop characteristics, state of erosion and relationship to Pleistocene glacial moraines. The glacial outwash gravels which extend south from the terminal moraine on Harrington Creek (Fig. 2) are of particular significance because they provide time control for some of the younger faulting and deformation in the area.

Two large areas of pediment gravels east of Harrington Creek have been grouped with the glacial outwash gravels (Qg) even though they may not be directly related to glaciation. The gravels rest on an erosional surface which has been uplifted and deeply eroded and therefore predates the alluvial fans (Qoa) which are graded near the present drainage system. The pediment gravels contain abundant large boulders and are similar to the glacial outwash deposits, whereas the Quaternary-Tertiary (QTg) gravels have smaller cobbles and boulders and a larger sand and clay fraction.

STRUCTURE

Paleozoic Structures

The Paleozoic structures in the region are folds and thrust faults. Early folds have east-west axes and involve Devonian and older rocks (Kerr, 1962; Decker, 1962). These folds predate the thrust faults in the area (Decker, 1962).

The Ordovician Valmy Group rocks exposed in the Independence Mountains are part of an allochthon of eugeosynclinal rocks which was thrust over lower Paleozoic carbonate rocks of the miogeosyncline (Decker, 1962; Kerr, 1962). This thrust is exposed in the central part of the Independence Mountains where it is called the "Seetoya thrust" (Kerr, 1962) and to the north in the Bull Run Quadrangle as the "Trail Creek thrust" (Decker, 1962). This thrust and the lower-plate carbonate rocks are not exposed in the study area, but 279 m of carbonates were penetrated in the bottom of the Tuscarora 66-5 hole, below 536 m of Valmy quartzite (Fig. 4). If the carbonates are part of the autochthonous eastern facies, then the base of the quartzite is a thrust fault. The time of thrusting is uncertain, but the rocks involved indicate post-Devonian(?) to pre-Miocene limits (Decker, 1962; Kerr, 1962).

The second thrust in the study area is at the base of the Schoonover Formation. Fagan (1962) defined the base of the Schoonover Formation as a fault, and Miller and others (1981) interpreted this fault to be a thrust. The thrust is dated as post–Late Permian on the basis of fossils in the lower plate, and it is overlain by Miocene rocks (Miller and others, 1981).

The fault at the base of the Schoonover Formation is exposed as a high-angle, east-west–trending fault along Jack Creek. Mullion structures on the exposed fault surfaces along Jack Creek indicate that the last fault movement was strike-slip and east-west. Measured dips on the exposed fault surface are 48 to 85 degrees to the north (Fig. 2) and may be vertical at its westernmost exposure. The Schoonover fault as exposed along Jack Creek could be interpreted as a right-lateral fault. It may be a tear fault offsetting the Schoonover thrust mapped to the northeast by Miller and others (1981). However, the Schoonover Formation dips steeply to the north (Fig. 2), and post-thrusting deformation may have tilted both the formation and the thrust fault to the north. Dips to the south of the fault between Jack Creek and Marsh Creek are generally low and randomly oriented. The northern dip of the Valmy Group rocks on the south side of Jack Creek steepens near the Schoonover fault (Fig.

2). It would appear, therefore, that the hinge line of an east-west-striking monocline is approximately coincident with the Schoonover fault.

Tertiary Structures

The main Tertiary structural elements are (1) north-south-trending normal faults, (2) a graben in the center of the study area, (3) a narrow horst and volcanic vent on the west side of the study area, (4) northwest-trending faults, and (5) north-south-trending folds. These structures are the result of east-west extension and volcanism. The north-south normal faults are the dominant Tertiary structures in the northern end of Independence Valley (Fig. 2). Perhaps the most significant of the faults is the range boundary fault on the west side of the Independence Mountains. This major structure trends north-south along most of the range front, but trends N30° W from Bull Creek to Marsh Creek under the alluvium, then turns to N10°E north of Marsh Creek (Fig. 2). The graben on the west side of this fault forms the Independence Valley and is about 4 mi wide in the center of the study area; it is bounded on the west by faults trending N10°E and N40° W. Drill-hole data provide some information on the magnitude of graben subsidence. In Tuscarora 66-5, the top of the Paleozoic rocks was encountered at an elevation of 1,280 m (Fig. 4), which is 670 m below their average elevation of 1,950 m along the range front. The east side of the graben appears to have subsided even more. Surface geology and the presence of tuffaceous sediments at a depth of 950 m (842 m elevation) in Tuscarora 51-9 indicate a graben between Hot Creek and Harrington Creek with subsidence of at least 1,300 m relative to the Paleozoic rocks to the east (Fig. 3).

A major structure, trending about north-south, is present along Hot Creek (Fig. 2). This structure is poorly exposed but has controlled emplacement of several basaltic-andesite plugs and the surface expression of the geothermal system. The close alignment of the intrusive plugs suggests that they may coalesce into a dike at depth. This dike is shown on cross sections B–B′ and C–C′ (Fig. 3). In section 30, T. 42 N., R. 52 E., on the east side of the horst, a fault-line collapse graben has formed (Fig. 2). This fault-produced graben is only slightly modified by erosion. A similar but more poorly defined graben is present a mile to the east in section 29.

The tuff-breccia vent is exposed on a horst on the west side of the study area. The horst extends north to Lime Mountain and is bounded by faults trending N10° E and N40° W (Fig. 2). The vent area has been uplifted relative to the rest of the horst, and the bounding faults of the vent, where well exposed, are convex upward. This evidence and the massive quartz veins within the vent and along some bounding faults suggest that an intrusion at depth may have uplifted the tuff-breccia (Fig. 3, section B–B′). The top of the tuff-breccia in Tuscarora 66-5 is about 600 m lower than the top of the eroded outcrop in the horst. Part of this may be due to original relief. The outcrop of Paleozoic rocks along Skull Creek, within the horst, is 490 m above the Paleozoic rock in Tuscarora 66-5.

Subordinate to the north-south structures are N20° W to N40° W faults. As mentioned above, a set of N30° W faults offset the range boundary fault between Bull Creek and Marsh Creek

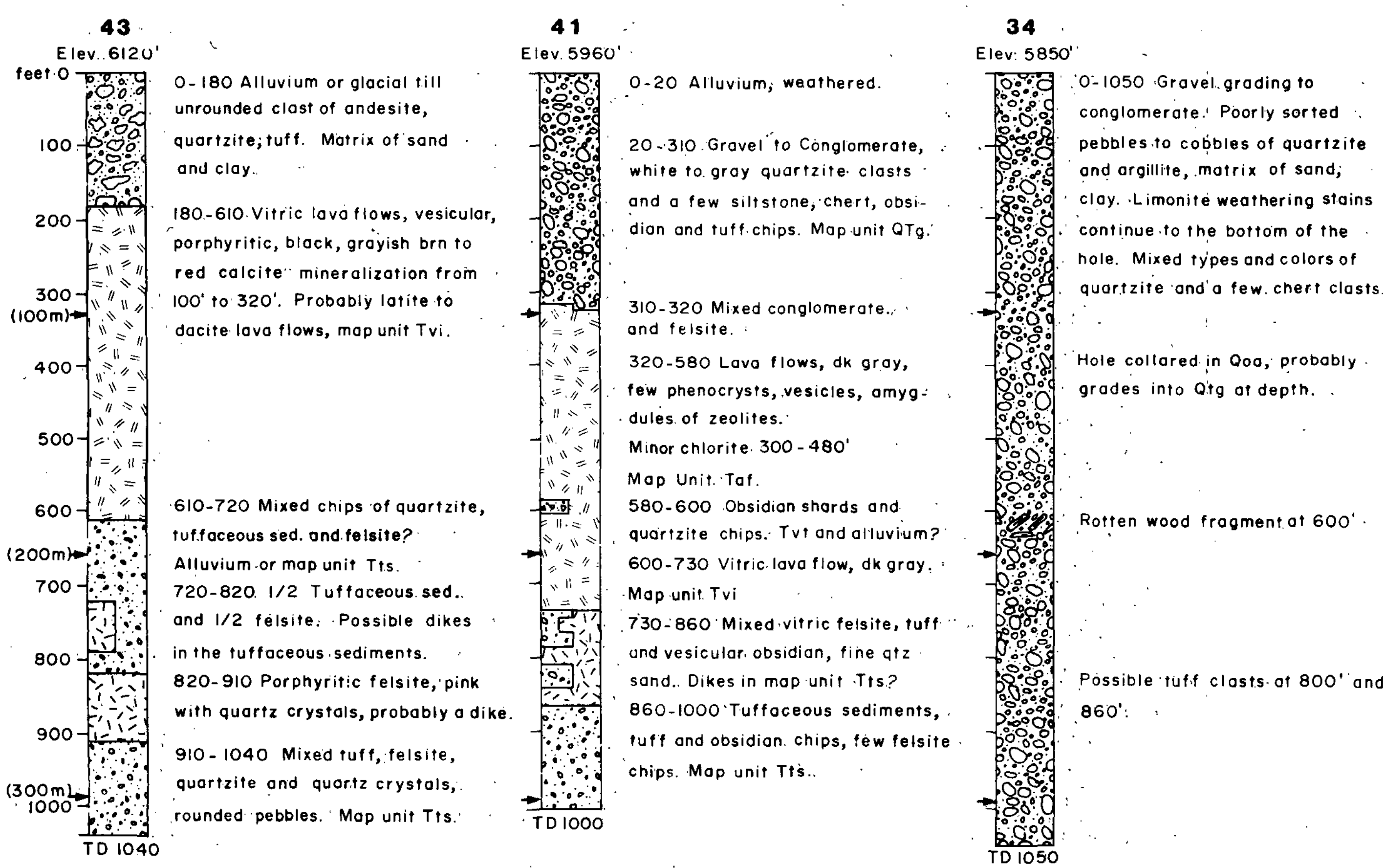

Figure 5. Lithologic logs of three thermal gradient holes. Unit intervals are given in feet.

(Fig. 2). This trend is continued into the Tertiary rocks; to the northwest, a N40° W fault offsets the east boundary of the horst (Fig. 2). The third and least developed fault trend is N30° E to N45° E. Faults of this trend extend northeast from the east side of the horst and also occur in the foothills of the Independence Mountains (Fig. 2).

An apparent stratigraphic reversal was encountered in Tuscarora 51-9, 1 km east of the hot springs (John Deymonaz, AMAX, 1981, oral commun.). A normal section, with the addition of an andesite intrusion, was penetrated in the upper 722 m of the hole, then bedded tuffaceous sedimentary rocks (Tts) were encountered below a 119-m-thick section of Paleozoic siltstone and shale (Fig. 3, C–C′). A reverse fault seems unlikely in Tertiary rocks when the dominant tectonic force was Basin and Range extension. Furthermore, a reverse fault, which would require more than 300 m of vertical displacement, does not fit with the surface geology. The simplest and most likely explanation for the lithologies encountered is that the drill hole drifted laterally across a high-angle fault, passing from Paleozoic rocks in the footwall to Tertiary rocks in the hanging wall (Fig. 3, C–C′). The cuttings of tuffaceous sedimentary rocks from the bottom of the hole were different from the tuffaceous sedimentary rocks in the upper part of the hole and therefore did not appear to be caving from the upper section. The hole was lost before a drift survey could be run.

Folding

Several folds, all with north-south axes, are present in the Tertiary rocks (Fig. 2). The glacial outwash gravels (Qg) in sections 3 and 10 along Harrington Creek have been deformed into a broad south-plunging syncline. The gravel deposit clearly originates from the terminal moraine at its north end and is therefore Pleistocene in age. The southern component of tilt may be a depositional slope, but the east-west downwarp of the uneroded surface has been produced by deformation. The syncline was probably produced by sag of the underlying tuffaceous sediments along the range boundary fault. On the other hand, the anticlines may have formed as a result of the tuffaceous sediments draping over small horsts in the Paleozoic basement rock. Another possible mechanism is slump folding due to the tuffaceous sediments sliding down the east-tilted fault block. Although the dips are steeper in the Tertiary units, the folding has affected the Pleistocene deposits as well.

A belt of gravel deposits lapping onto the Tertiary rocks extends along the northwest side of the recent alluvial fan formed by Harrington Creek (Fig. 2). Although bedding orientation is not easily determined in the field due to the coarse boulders and unconsolidated nature of the gravels, the outcrop pattern and photo analysis suggest a southeast dip. This is compatible with the interpretation that the syncline in sections 3 and 10 continues to the southwest and broadens. Harrington Creek's alluvial fan has filled this large structural depression. This interpretation would suggest a considerable thickness of Quaternary alluvium in the north end of Independence Valley and a major northeast-trending fault which controlled formation of the syncline. Some evidence of the thick alluvium is given by thermal gradient hole 34 drilled in the SW1/4SW1/4, sec. 14, T. 41 N., R. 52 E. (Fig. 2). This hole penetrated 317 m of alluvial gravels and conglomerate without encountering Tertiary volcanic rocks or Paleozoic rocks (Fig. 5). Interpretation of dipole-dipole resistivity data across the Harrington Creek alluvial fan indicates a thick section of conductive materials which could be water-saturated alluvium (C. E. Mackelprang, 1980, personal commun.).

Thermal Phenomena

Numerous hot springs and extensive sinter deposits occur along Hot Creek (Fig. 6). The main sinter mound is 1,000 m by 330 m and 35 m high, is covered with sagebrush, and has no active springs on it. The siliceous sinter mound extends down a west-facing slope from several faults, and the sinter thickness at any one point probably does not exceed 10 m. At the west foot of the mound, three springs in the alluvium are currently depositing silica (Fig. 6).

Most of the active springs occur in a small area 400 m upstream from the large sinter mound. The springs form a roughly triangular pattern, and spring temperatures are 55 to 95 °C. The hotter springs issue directly from tuffaceous sediments exposed in Hot Creek's active channel. These springs are depositing both siliceous and calcareous sinter, sulfur, and sublimates. A few of the springs are boiling at the surface, and there is steam from a small vent, probably supplied by near-surface boiling.

The Na-K-Ca geothermometer indicates a reservoir temperature of 181 to 228 °C. A Cl-SiO_2-enthalpy mixing model indicates an equilibrium temperature for the reservoir of 216 °C (Pilkington and others, 1980). The thermal waters are a sodium-bicarbonate type, and the reservoir temperature of Hot Sulphur Springs was calculated to be 183 °C by the silica conductance method (Mariner and others, 1974).

The mapped travertine deposit in Hot Creek consists of 0.3 to 1 m thick, coarse-grained, white calcite blocks which are currently being eroded. The calcite blocks lie on multi-colored, hematite-stained clay which is probably altered tuffaceous sediments.

Hot Creek is a perennial cold-water creek above Hot Sulphur Springs. The stream water is warm below the hot springs due to the addition of the thermal water. Because of the intermixing of stream and thermal waters in the thin alluvium, the volume of water discharged by the thermal springs alone cannot be estimated.

A few small thermal springs occur to the south and southwest of the main thermal area. A thermal spring on the west side of Hot Creek (Fig. 6) issues from the top of a low calcite and sediment mound. Two siliceous sinter mounds about 40 cm high are present on the calcareous deposit. An intermittently flowing thermal spring occurs 900 m south of the main sinter mound (Fig. 6), and a spring with a flow of 75 to 100 l/min, 3 km south-southwest of the main sinter mound, has a temperature of 21 °C as compared to 10 °C for cold springs in the area. Petaini Springs, near the mouth of Jerritt Canyon, 11 km southeast of Hot Sulphur Springs, is the only other spring in the valley reported to be warm (Garside and Schilling, 1979).

The subsurface structures and lithologies controlling the movement of the thermal fluids and the location of the reservoir are unknown. Dipole-dipole resistivity data (Berkman, 1981; Mackelprang and others, 1981) indicate a deep conductive zone under the valley to the southeast of the springs. This conductive zone approaches the surface at the hot springs. This conductive zone is about 1,200 m under the valley and may represent the geothermal system.

CONCLUSIONS

A thick accumulation of Oligocene tuffaceous sediments and tuffs overlain by Miocene lava flows has filled a deep graben bounded by faults trending north to N10° E. Subsidence has been greatest (over 1,200 m) along a 1.5 to 3 km wide trough, bounded on the east side by the range frontal fault of the Independence

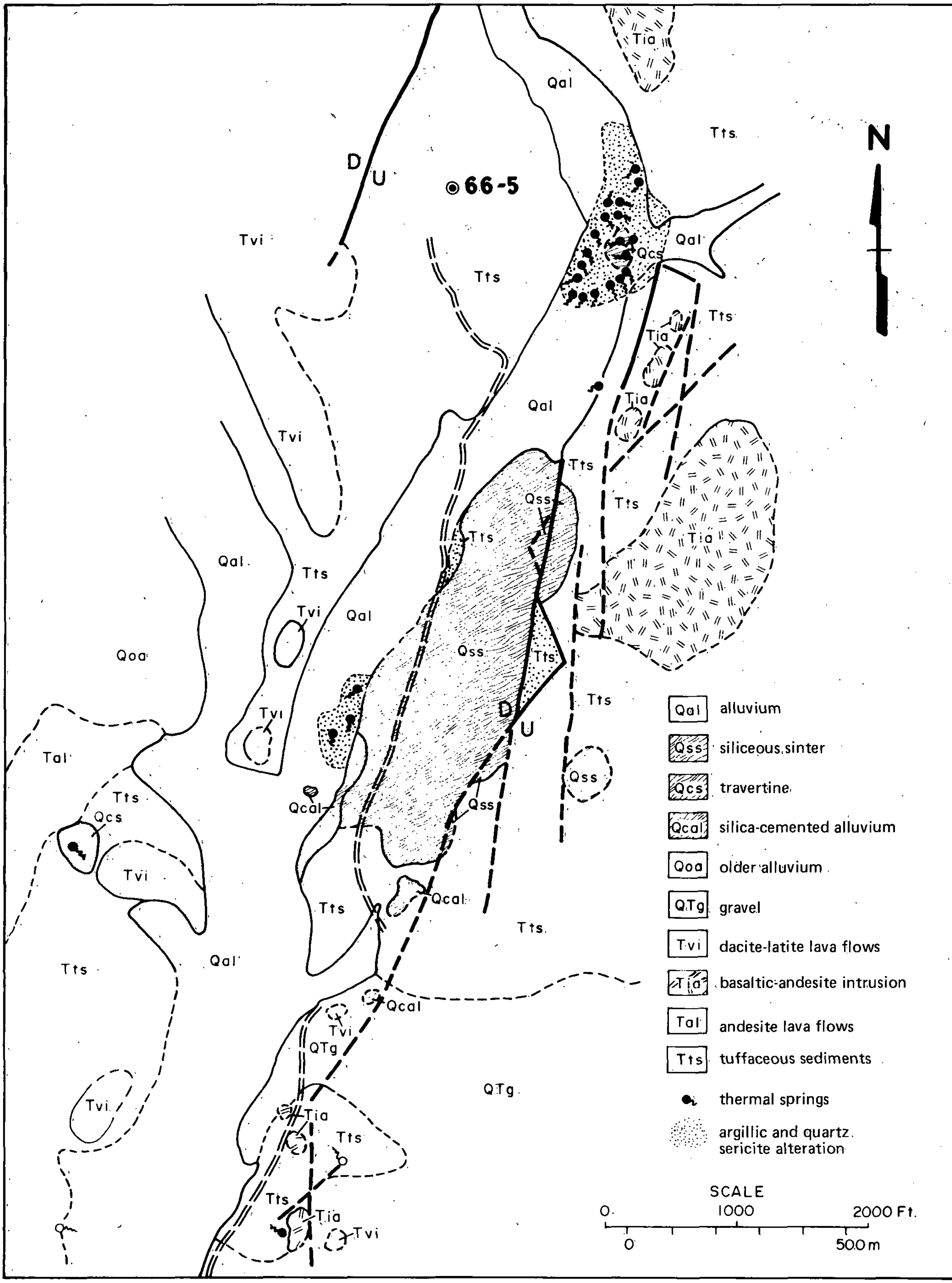

Figure 6. Detailed geologic map of Hot Sulphur Springs. See Figure 2 (foldout) for location.

Mountains and broadening to the south. Late Tertiary deformation which has continued into the Quaternary has produced north-south–trending synclines and anticlines and faults trending north-south and N20° W. The folds have probably resulted from fault sag and slump folding.

The surface expression of the Hot Sulphur Springs thermal system is controlled by a fault zone trending N20° E. Exposed argillic alteration produced by the thermal system is limited to the spring area. Quartz-sericite alteration which predates the present thermal system is present along the fault zone.

The subsurface character of the geothermal system is not known, but the geophysical and geologic data are consistent with an interpretation that the reservoir is 3 to 5 km southeast of the hot springs. In this model, meteoric water circulates down along the range front fault system and is heated at depth. The thermal waters rise along major fractures, perhaps the intersection of the N10° E and N30° W fault zones, into either a solution reservoir in the lower-plate carbonates or a fracture reservoir in the overlying Valmy Group quartzite. The fracture reservoir and feeder channel ways may have been formed by brecciation along the thrust fault and by formation of the deep graben. The reservoir cap consists of the incompetent and less permeable Tertiary tuffs and tuffaceous sedimentary rocks, the base being 1,200 m or more below the surface. Some of the thermal fluids migrate up major fractures within the Paleozoic shale, chert, and greenstone unit which overlies the Valmy Group quartzite. The fluids probably move updip to the northwest along gravel aquifers either at the base of or within the tuffaceous sedimentary rocks, ultimately reaching the surface along the faults at the hot springs. Cold water aquifers in the thick quartzite gravel overlying the tuffaceous sedimentary rocks apparently mask the thermal anomaly directly above the reservoir.

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REFERENCES CITED

Berkman, F. E., 1981, The Tuscarora, Nevada geothermal prospect: A continuing case history [abs.]: Geophysics, v. 46, no. 4, p. 455–456.

Churkin, M., Jr., and Kay, M., 1967, Graptolite-bearing Ordovician siliceous and volcanic rocks, northern Independence Range, Nevada: Geological Society of America Bulletin, v. 78, p. 651–668.

Decker, R. W., 1962, Geology of the Bull Run Quadrangle, Elko County, Nevada: Nevada Bureau of Mines Bulletin 60, 65 p.

Fagan, J. J., 1962, Carboniferous cherts, turbidites, and volcanic rocks in northern Independence Range, Nevada: Geological Society of America Bulletin, v. 73, p. 595–611.

Garside, L. J., and Schilling, J. H., 1979, Thermal waters of Nevada: Nevada Bureau of Mines and Geology Bulletin 91, 160 p.

Hope, R. A., and Coats, R. R., 1976, Preliminary geologic map of Elko County, Nevada: U.S. Geological Survey Open-File Report 76-779, scale 1:100,000.

Kerr, J. W., 1962, Paleozoic sequence and thrust slices of the Seetoya Mountains, Independence Range, Elko County, Nevada: Geological Society of America Bulletin, v. 73, p. 439–460.

Mackelprang, C. E., Lange, A. L., Sibbett, B. S., and Pilkington, H. D., 1981, Interpretation of a telluric-magnetotelluric survey at the Tuscarora geothermal exploration unit, Elko County, Nevada, *in* Technical Papers, v. 1: Society of Exploration Geophysicists, Annual International Meeting and Exposition, 51st, p. 205–226.

Mariner, R. H., Rapp, J. B., Willey, L. M., and Presser, T. S., 1974, Chemical composition and estimated minimum thermal reservoir temperatures of the principal hot springs of northern and central Nevada: U.S. Geological Survey Open-File Report 74-1066, 30 p.

McKee, E. H., Tarshis, A. L., and Narvin, R. F., 1976, Summary of radiometric ages of Tertiary volcanic and selected plutonic rocks in Nevada. Part V: Northeastern Nevada: Isochron/West, no. 16, p. 15–27.

Miller, E., Bateson, J., Dinter, D., Dyer, R., Harbaugh, D., and Jones, D. L., 1981, Thrust emplacement of Schoonover Sequence, Northern Independence Mountains, Nevada: Geological Society of America Bulletin, v. 92, p. 730–737.

Pilkington, H. D., Lange, A. L., and Berkman, F. E., 1980, Geothermal exploration at the Tuscarora Prospect in Elko County, Nevada: Geothermal Resources Council Transactions, v. 4, p. 233–236.

Roberts, R. J., Hotz, R. E., Gilluly, J., and Ferguson, H. G., 1958, Paleozoic rocks of north-central Nevada: American Association of Petroleum Geologists Bulletin, v. 42, p. 2813–2857.

Schilling, J. H., 1965, Isotopic age determinations of Nevada rocks: Nevada Bureau of Mines Report 10, 79 p.

Stewart, J. H., and Carlson, J. E., 1976, Cenozoic rocks of Nevada: Nevada Bureau of Mines and Geology, Map 52, 5 p., scale 1:100,000.

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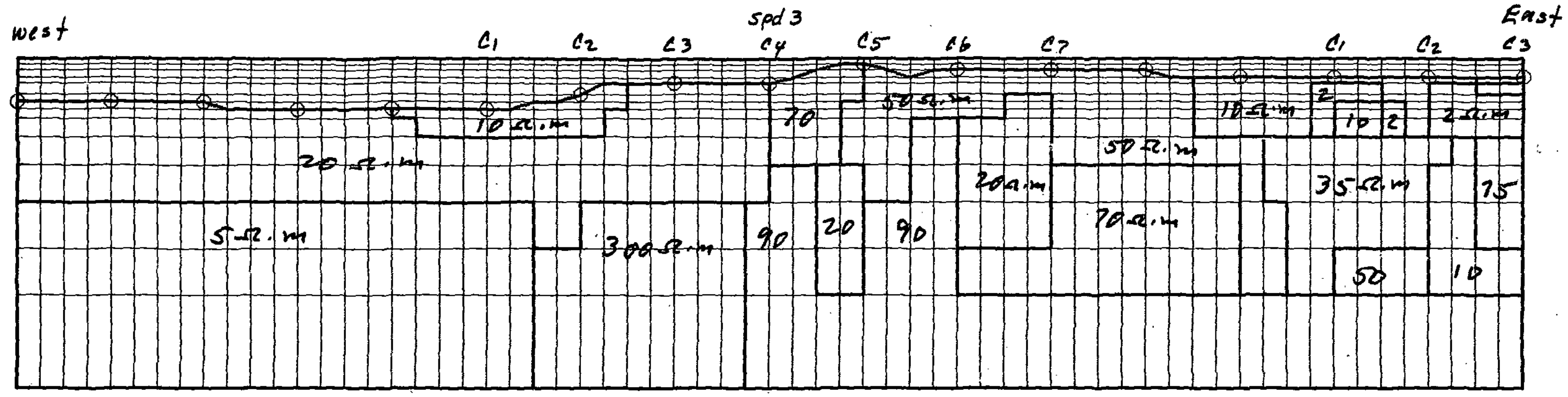

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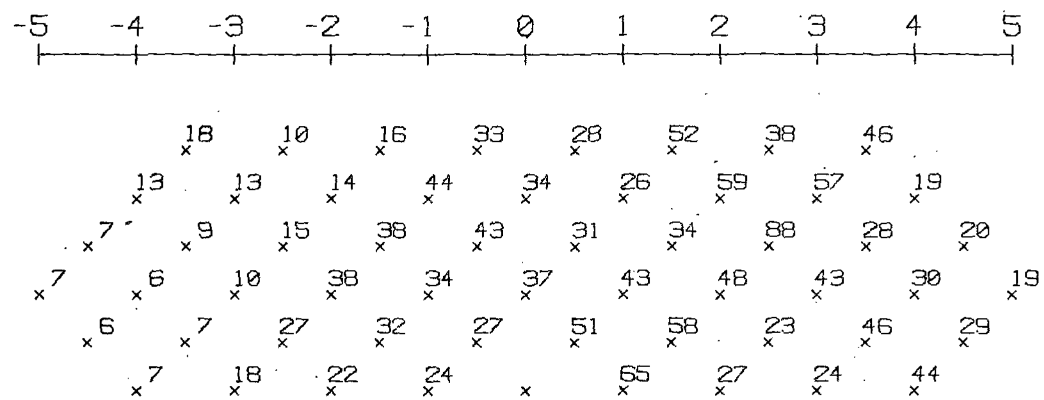

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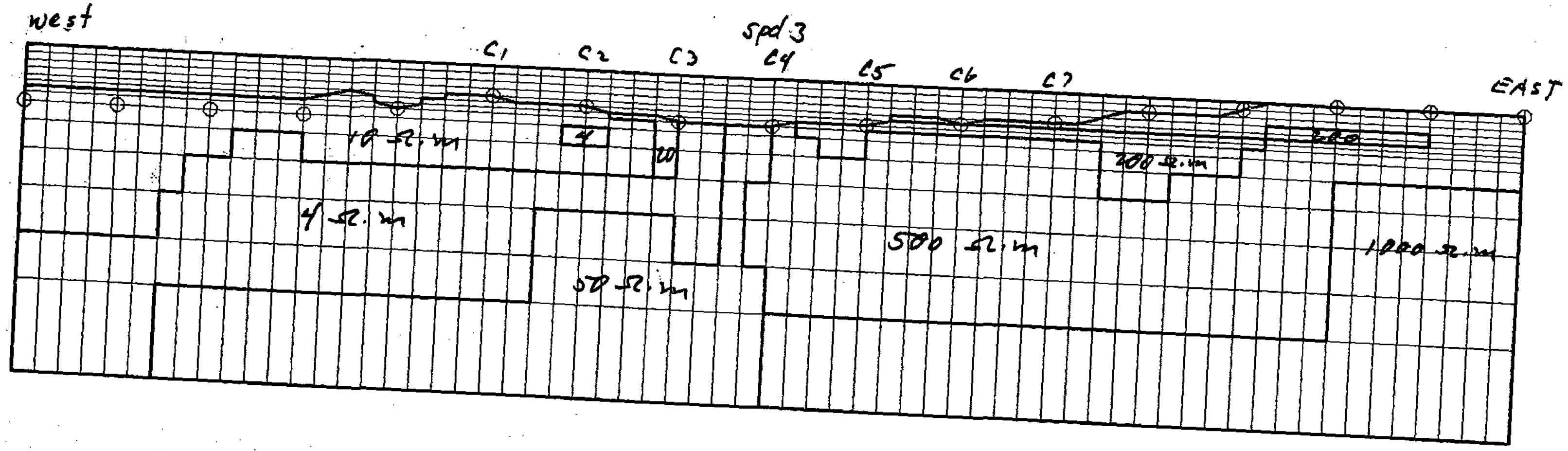

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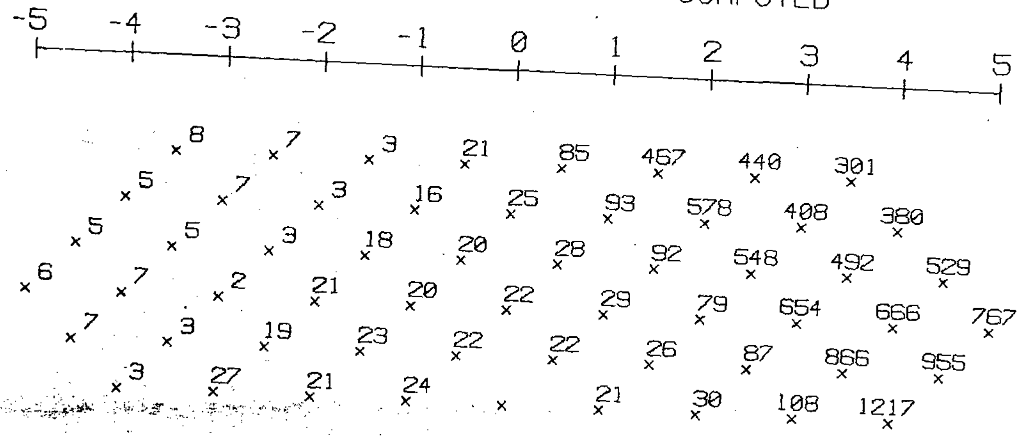

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Tuscarora Project - Elko Co., NEVADA
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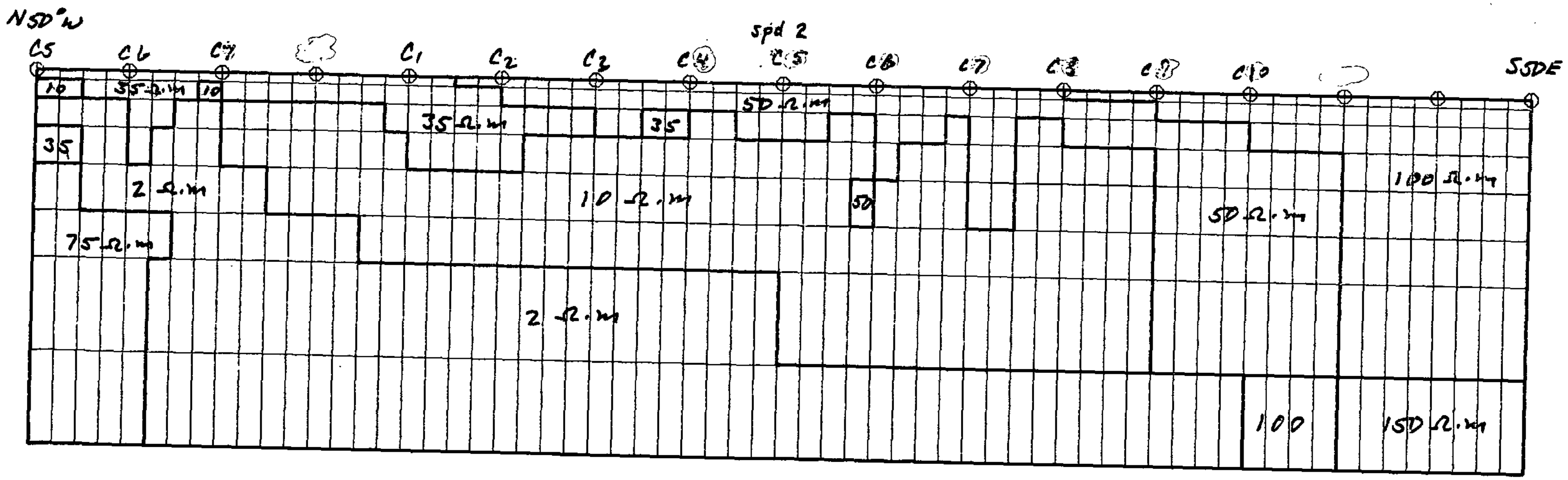

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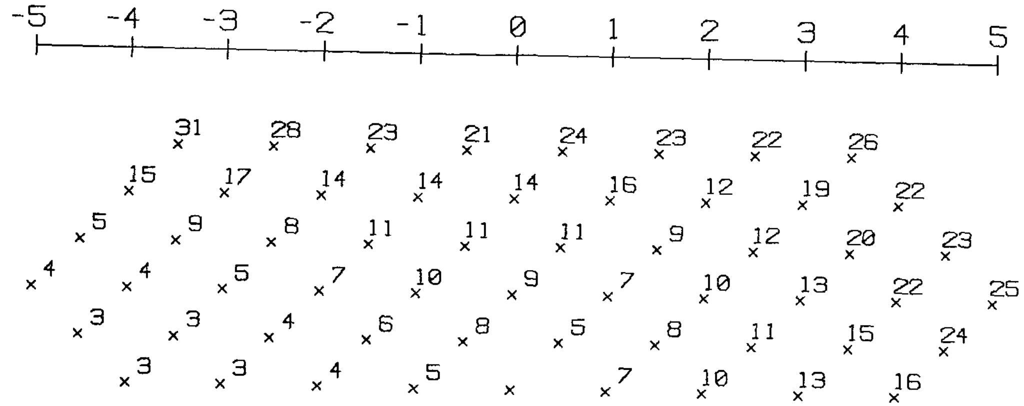

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Tuscarora Project - Elko Co., Nevada

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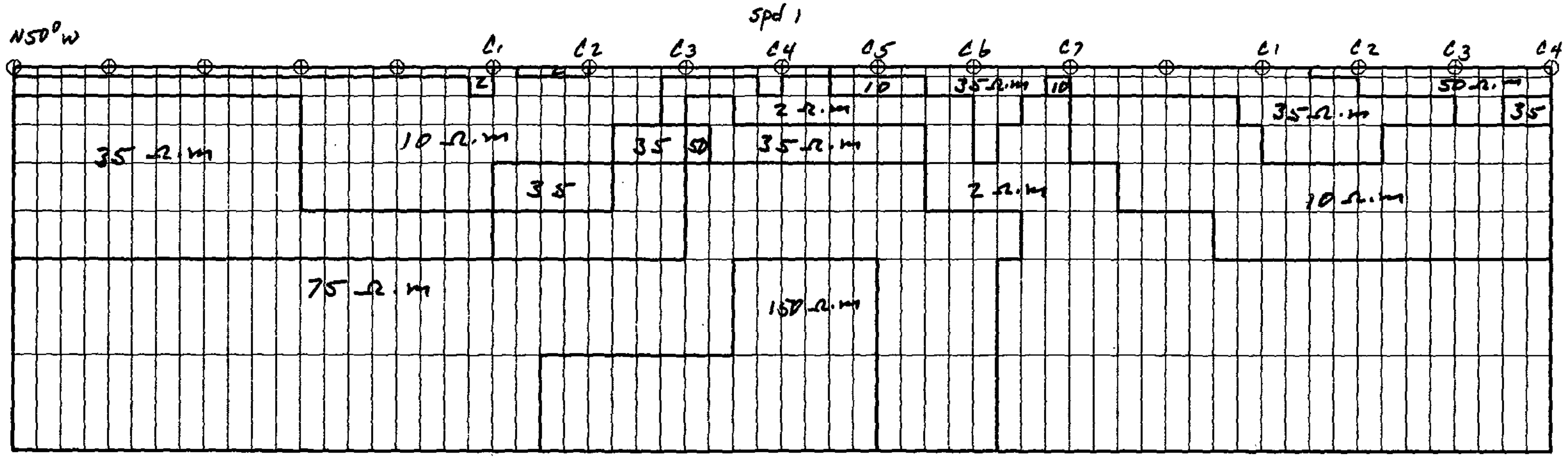

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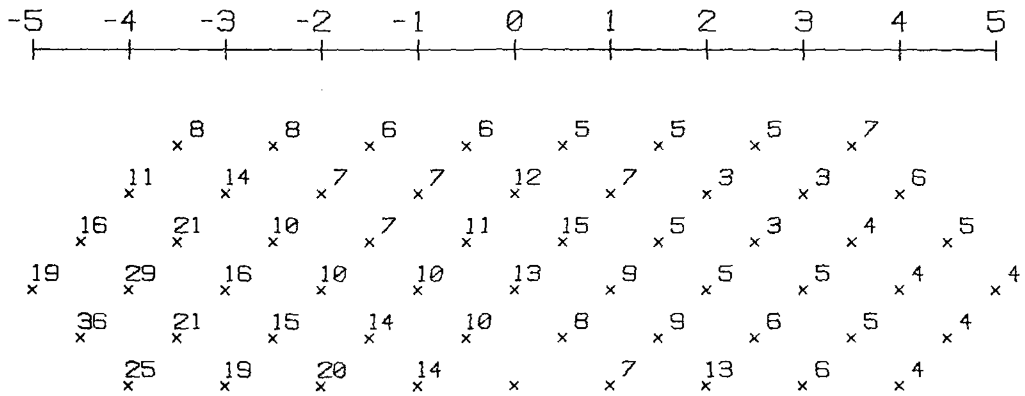

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Tuscarora Project - Elko Co., NEVADA

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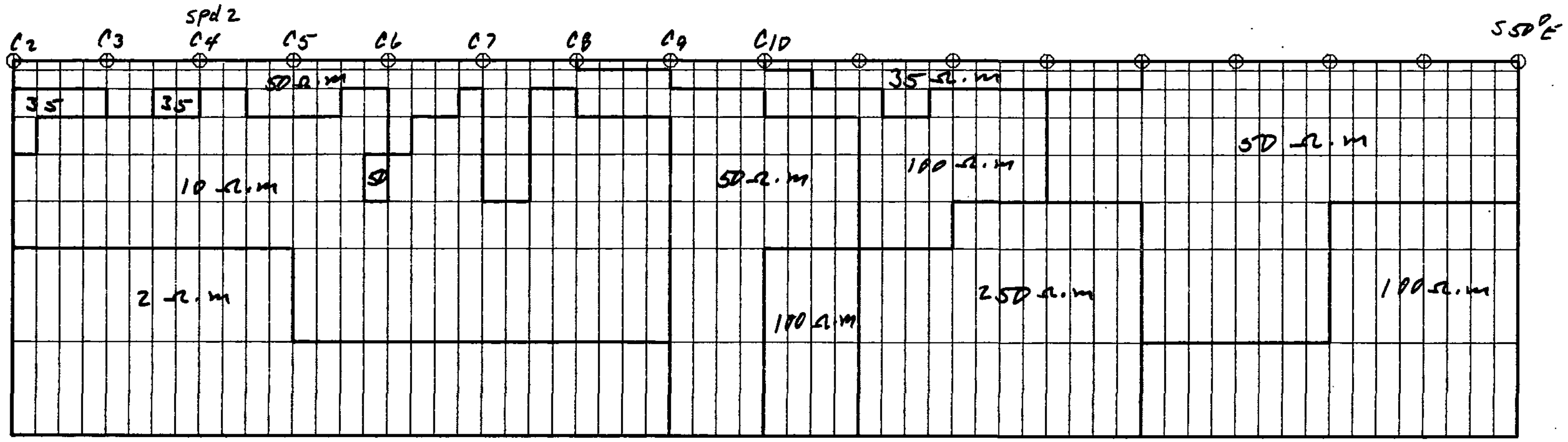

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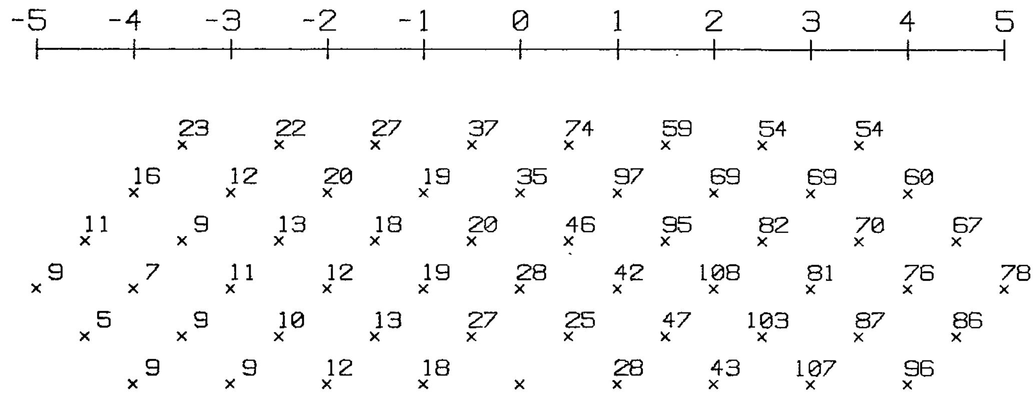

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Tuscarora Project - Elko Co., Nevada

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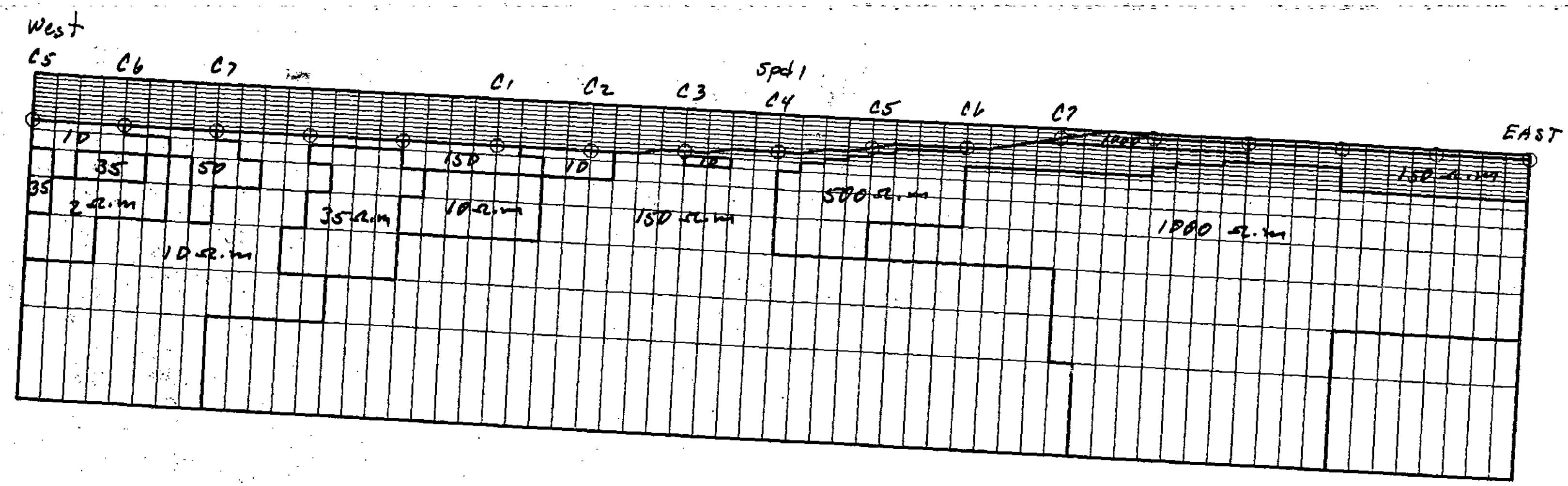

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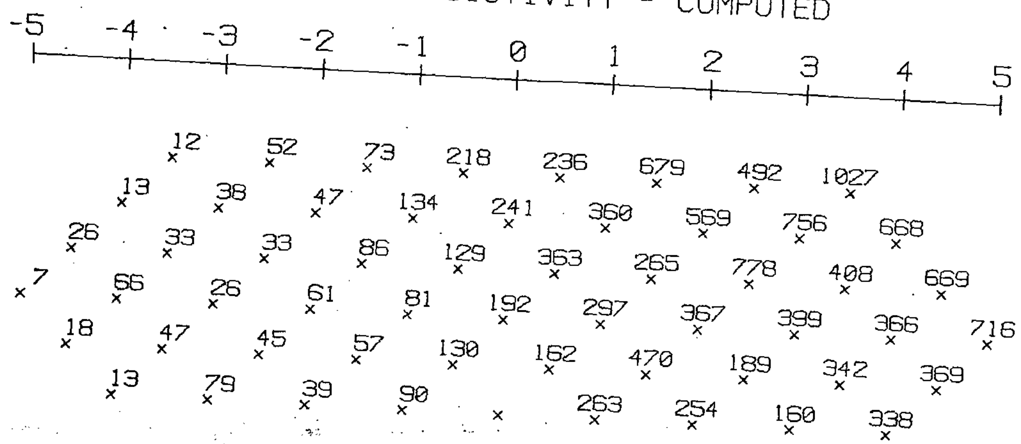

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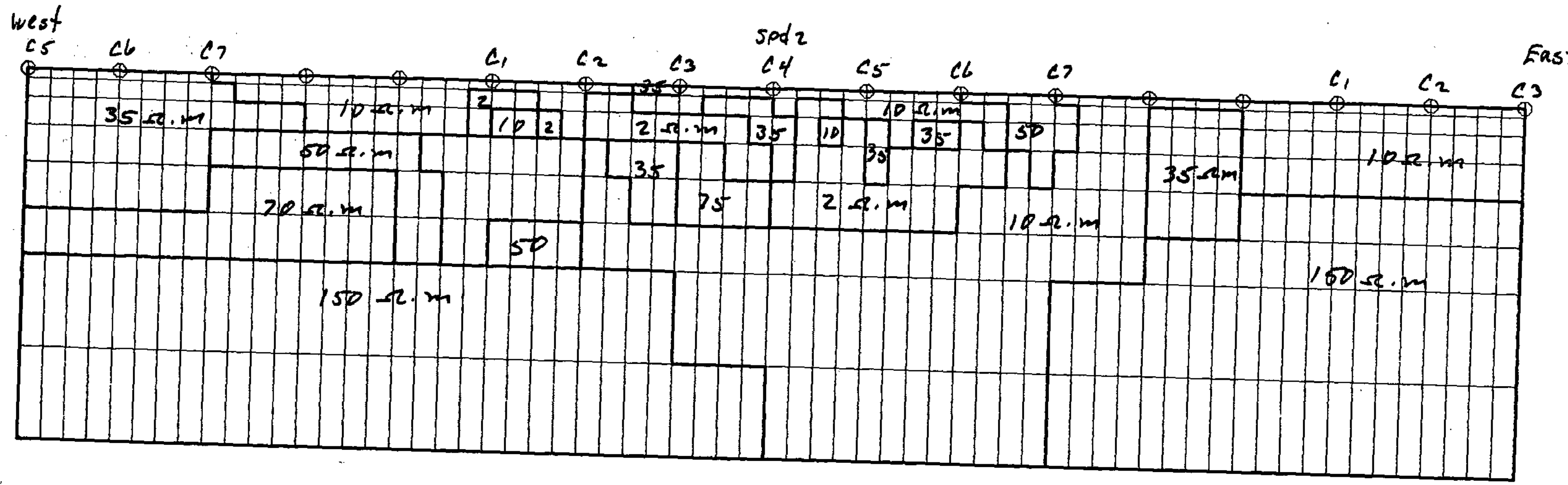

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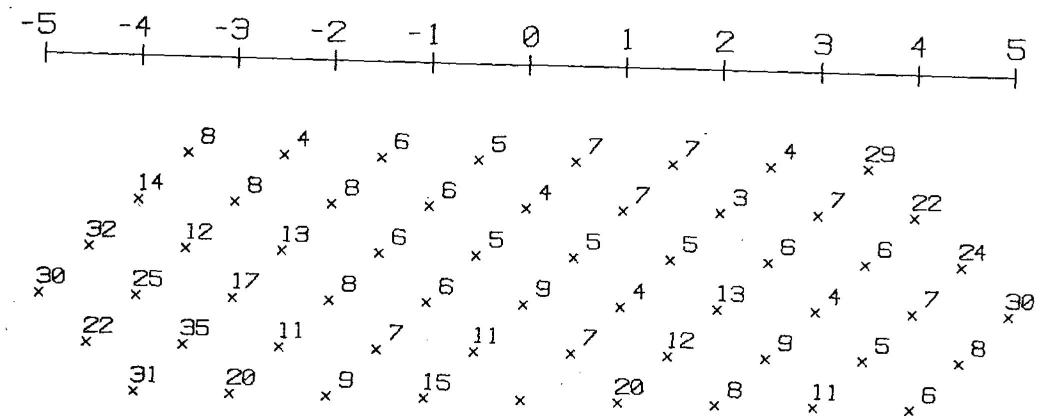

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