

ISSN number 0971 - 9709



IGU
Promoting Earth System Sciences

The Journal of Indian Geophysical Union

A SCI Journal

Impact Factor (2026) 0.3



AN OPEN ACCESS BIMONTHLY JOURNAL OF IGU

VOLUME 30, ISSUE 5, SEPTEMBER, 2026

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The Journal with six issues in a year publishes articles covering
Solid Earth Geosciences; Marine Geosciences; and Atmospheric, Space and Planetary Sciences.
The Journal is supported by CSIR-NGRI & MoES, Govt. of India

Annual Subscription

Individual Rs -1000/- per issue and Institutional Rs- 5000/- for six issues
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 ([https://doi.org/10.71122/JIGU.30\(4\)2026.0024](https://doi.org/10.71122/JIGU.30(4)2026.0024)).
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Crustal reworking and magmatic evolution in the northeastern Dharwar craton: evidence from geochemistry and zircon U-Pb geochronology of Ghanapur granitoids

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ABSTRACT

The widespread occurrence of granitoids in Precambrian terranes provides crucial insights into crustal processes and geochemical evolution within cratonic blocks. The Eastern Dharwar Craton (EDC) of southern India, records extensive granitic magmatism during the late Archean, a transformative period in Earth's history marked by significant geological and geochemical shifts. This study presents a comprehensive, first-hand investigation of granitoids from the Ghanapur area in the underexplored north-eastern EDC, using integrated fieldwork, petrography, geochemistry, and zircon U-Pb geochronology, to provide new insights into the evolutionary history of granitic crust. The granitoids of Ghanapur are categorized into three suites: granitic gneisses, granodiorite gneisses and leucogranites. Zircon U-Pb dating reveals two distinct magmatic episodes: granitic gneisses at ca. 2697–2607 Ma, while granodiorite gneisses formed at ca. 2560–2545 Ma. Granitic gneisses are silica rich, weakly metaluminous to slightly peraluminous, and show high-K calc-alkaline to shoshonitic affinity, with magnesian character. They show variable Eu anomaly and REE fractionation. Their ages and geochemistry together suggest repeated crustal reworking and mixed origin. The granodioritic gneisses are weakly peraluminous and moderately silicic, with elevated Mg# and enriched in incompatible elements. They resemble Archean sanukitoids and suggest their origin from a metasomatized mantle wedge. These rocks record a two-stage evolutionary history: initial sanukitoid magmatism driven by partial melting of the enriched mantle wedge, followed by a younger migmatitic episode, resulting in their preservation as mesosome within the migmatitic complex. Leucogranites are garnet- or magnetite-bearing, silica-rich, peraluminous to strongly peraluminous, high-K calc-alkaline, ferroan rocks, likely reflecting the waning stages of late Archean crustal thickening. Taken together, the geochemistry and geochronology suggest progressive crustal reworking and differentiation over a period of at least 150 Ma during the late Archean in the Ghanapur region.

Keywords: Ghanapur granitoids, NE Dharwar craton, U-Pb Zircon dating, Sanukitoid, Crustal evolution, Leucogranites, Gneisses

INTRODUCTION

The late Archean (2.8–2.5 Ga) was a transformative period in Earth's history, marked by rapid continental growth and fundamental changes in tectonic style (Hawkesworth et al., 2020). While the precise volume of crust generated remains debated (e.g., Dhuime et al., 2012; Rosas and Korenaga, 2018), there is a broad consensus that this interval witnessed a compositional shift from sodic tonalite-trondhjemite-granodiorite (TTG) suites toward more potassic granitoids (Moyen, 2011; Laurent et al., 2014). The late Archean potassic granitoids are petrogenetically diverse, encompassing sanukitoids, two-mica granites, biotite granites, and hybrid varieties, each recording different source contributions and melt-generation processes (Laurent et al., 2014). Their appearance in the late Archean record is commonly linked to the progressive evolution of source compositions (Johnson et al. 2019) together with changing mantle thermal conditions and evolving subduction dynamics (Dash et al., 2025). Understanding their petrogenesis and relative timing provides important constraints on how Archean crust matured from juvenile additions to stabilized cratonic lithosphere.

The Eastern Dharwar Craton (EDC) of southern India, preserves an excellent record of the late Archean magmatism. The EDC is considered to have been accreted mostly during

the late Neoarchean (2.7–2.5 Ga) (Peucat et al., 1993; Jayananda et al., 2000, 2013), and is characterized by numerous NW-SE trending linear volcano-sedimentary sequences of Neoarchean with volumetrically dominant 2.57–2.50 Ga K-rich granitoids, and 2.67–2.58 Ga transitional TTGs (Jayananda et al., 2013, 2018; Dey et al., 2017). Various tectonic models, such as subduction-driven accretion, mantle plume-derived crustal growth, and accretion via combined plum-arc processes are used to explain crustal accretion in EDC (Chadwick et al., 2000; Manikyamba and Kerrich, 2012; Dey et al., 2017; Jayananda et al., 2018).

However, the current understanding of crust formation and tectonic evolution of EDC is mostly restricted to the southern and western parts of the craton whereas the northern part of the craton largely remains underexplored with limited comprehensive geochemical and geochronological studies. This study presents field, petrographic, geochemical, and zircon U-Pb geochronological data from granitoids of the Ghanapur area in the north-eastern EDC. We document three granitoid suites, granitic gneiss, granodiorite gneiss and leucogranite and constrain their petrogenesis, age relationships, and tectonic significance. The results provide new insights into late Archean sanukitoid magmatism, crustal anatexis, and the role of migmatization in crustal reworking within this underexplored segment of the craton.

GEOLOGICAL SETTING

The Dharwar craton is a classic Archean granite–greenstone terrane of southern India and is conventionally divided into the Western and Eastern Dharwar cratons on the basis of lithology, metamorphic grade, and crustal evolution (Chadwick et al., 2000; Jayananda et al., 2006; Chardon et al., 2008) (Figure 1a). The EDC is dominated by Neoproterozoic TTG gneisses, younger potassic granitoids, migmatites, and linear supracrustal belts (Peucat et al., 1993; Jayananda et al., 2000). It also preserves a wide range of late Neoproterozoic granitoids, including TTG, high-Mg diorite/sanukitoid suites, Closepet-type granites, and K-rich anatectic leucogranites, along with supracrustal assemblages recording two main age populations at ca. 2.75–2.67 Ga and 2.58–2.54 Ga (Balakrishnan et al., 1999; Moyen et al., 2003; Rogers et al., 2007; Manikyamba and Kerrich, 2012; Jayananda et al., 2013, 2018; Dey et al., 2017; Bhaskar Rao et al., 2020; Ashok et al., 2021). Craton-wide magmatism at ca. 2.56–2.52 Ga, marked by high-K calc-alkaline granitoids, represents the final stage of crustal stabilization in the EDC (Anand et al., 2014; Dey et al., 2017; Jayananda et al., 2018).

The northern and northeastern EDC show marked lithological and metamorphic variation, ranging from lower-grade TTG–migmatite–granite–supracrustal terrains in the

south to high-grade granulite crust around Karimnagar in the north (Figure 1a). Available geochronological data from the northern EDC are limited and are largely confined to granulite-facies rocks, which record emplacement ages of ca. 2.8–2.5 Ga and metamorphic overprints around 2.6 Ga (Rajesham et al., 1993; Yoshida et al., 1996; Santosh et al., 2004; Sarbajna et al., 2013; Prakash et al., 2017).

Geology of the Ghanapur area

The Ghanapur area lies within low-grade granite greenstone terrain of north eastern EDC and is characterized by variably foliated granitoids, migmatitic gneisses, and associated supracrustal rocks exposed as discontinuous outcrops (Figure 1b). The variably foliated granitoids and the migmatitic gneisses represent different structural states of the same granitoid assemblage. The granitoids preserve a gneissic foliation of variable intensity and, affected by younger anatectic event. This gives appearance of migmatitic outcrop with granitic leucosome and granodioritic mesosome, without being genetically related with each other. The granitoids commonly show intrusive relationships with the associated metasedimentary and mafic units, and many outcrops preserve enclaves, stretched xenoliths, and quartzofeldspathic segregation bands aligned parallel to the regional foliation.

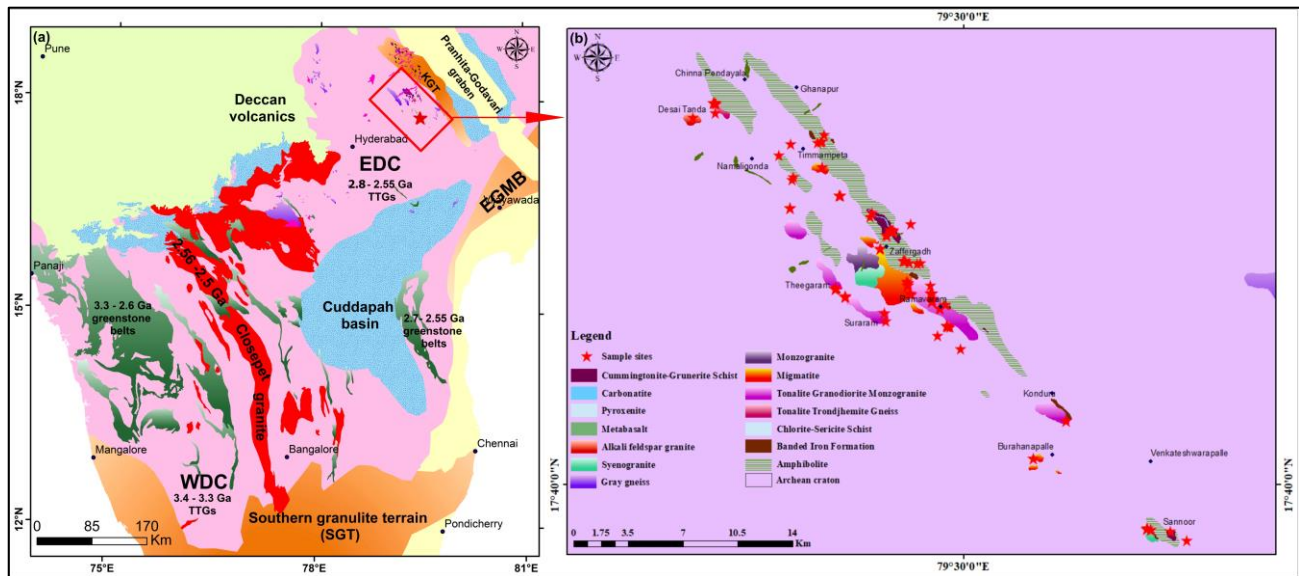


Figure 1. (a) A simplified geological map of southern India showing Archean Dharwar craton with dominant lithology. **(b)** Geological map of Ghanapur Belt (study area) with sampling locations, modified after the Geological Survey of India. WDC- Western Dharwar Craton, EDC- Eastern Dharwar Craton, EGMB- Eastern Ghat Mobile Belt, SGT- Southern Granulite Terrain, KGT- Karimnagar Granulite Terrain

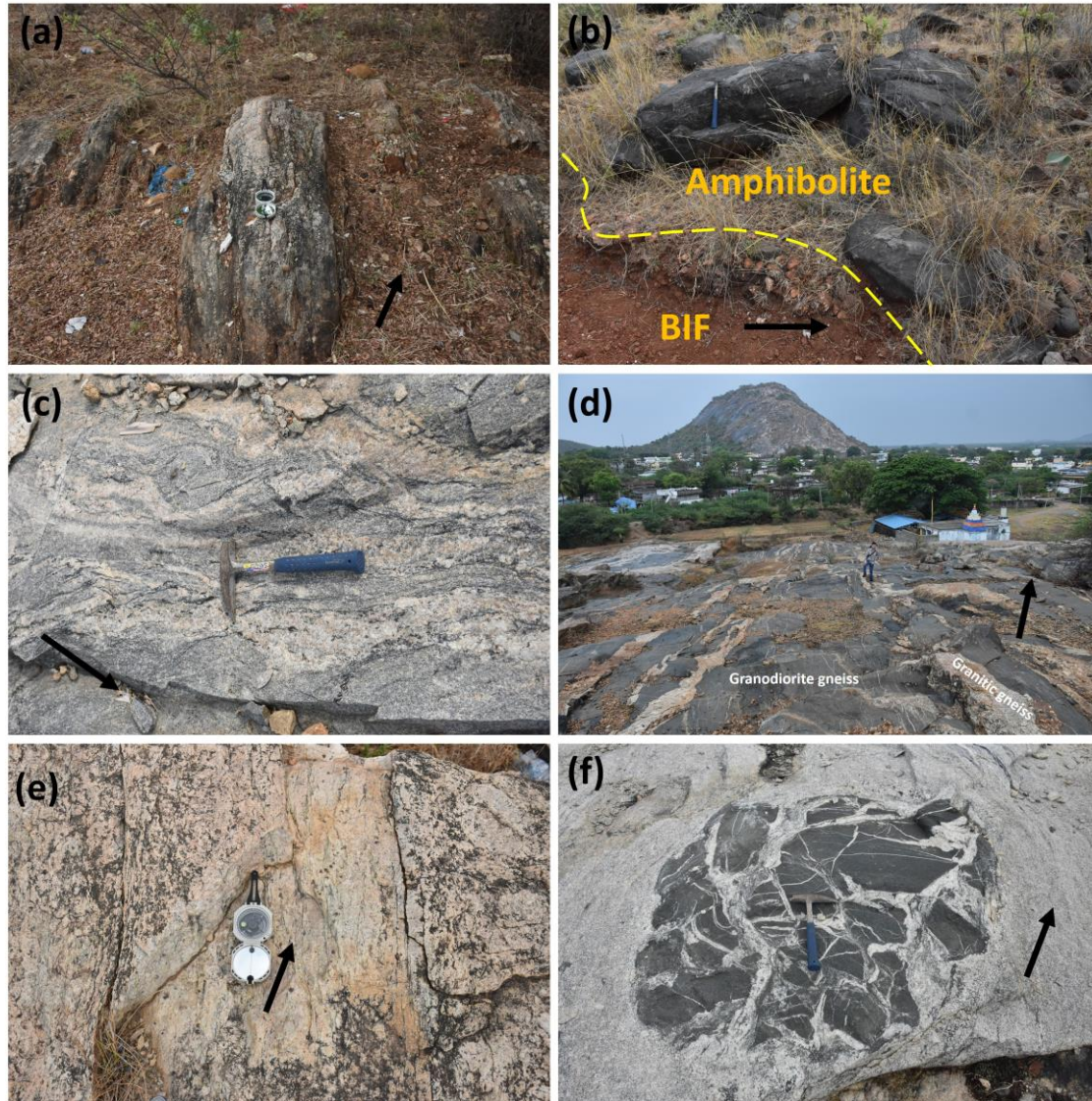


Figure 2. Field photographs with black arrow pointing north. **(a)** BIF exposure near Pamnoor village, showing alternate bands of magnetite and quartzite. **(b)** Contact of weathered BIF and amphibolite at foothill of Durgaya Tanda. **(c)** Coarse grained granitic gneiss from quarry of Theegaram-gutta. **(d)** Granodiorite gneiss of Zaffergadh hill showing variably deformed quartz veins with pygmatic folding intruded along the general gneissic foliation. **(e)** Outcrop of garnet bearing leucogranite near Durgaya Tanda. **(f)** Mafic enclave within granitic gneiss showing multiple injections of gneissic material.

In the Ghanapur area, the rock assemblage is dominated by granitoid gneisses with local amphibolites, schistose metasedimentary rocks, and banded iron formation (BIF). Field exposures of representative outcrops are given in Figure 2a-f. Earlier mapping by Guguloth and Manimaran (2019) recognized several granitoid varieties, including tonalite-trondhjemite gneiss, granodiorite to monzogranite, syenogranite, alkali feldspar granite, and garnet-bearing leucogranite. The granitoids are typically medium- to coarse-grained and locally porphyritic, with mafic enclaves and schlieren elongated parallel to the gneissic fabric. The associated metasedimentary and mafic units display compositional banding, tight folding, and localized shear

fabrics, while quartz veins intruded along the gneissic foliation are folded, boudinaged, and locally crosscut by younger quartz and epidote veins. Later brittle fractures and joints are also present in several litho-units. Together, these field relationships indicate a polyphase deformation history and provide the geological context for the present study.

METHODOLOGY

Sampling

This study focuses on three granitoid suites from representative outcrops in the Ghanapur, Zaffergadh, and Sannoor areas: granitic gneiss, granodiorite gneiss, and

leucogranite. Sample locations are shown in Figure 1b. Samples were selected based on field relationships, megascopic observations to represent the lithological diversity of the granitoid suite. Geochemical analyses were carried out on 39 samples, including 20 granitic gneisses, 12 granodioritic gneisses, and 7 leucogranites, while zircon U–Pb geochronology was undertaken on two granitic gneisses and two granodioritic gneisses.

Analytical techniques

Major and trace-element analyses of granitoid samples were carried out at National Centre for Earth Science Studies (NCESS), Trivandrum, and CSIR-National Geophysical Research Institute (NGRI), Hyderabad, respectively. Before analysis, the samples were cleaned, crushed, ultrasonically washed, and powdered to about 200 mesh. Loss on ignition (LOI) was determined by heating 1 g of powdered sample at 950 °C for 30 min.

Major elements were analyzed on fused glass beads prepared with 1 g of sample powder and 5 g of lithium borate flux using a Katanax X-300 X-Fluxer triple-station fusion apparatus, followed by measurement on a Bruker S8 Tiger sequential wavelength-dispersive X-ray spectrometer. Trace elements were determined by digesting 50 mg of powdered sample in HF–HNO₃–HClO₄ (3:1:0.5) in Savillex® Teflon beakers at 70 °C for 5–7 days, followed by evaporation and dissolution in 2% HNO₃. The solutions were analyzed using an Agilent 7800 quadrupole ICP-MS with Rh as the internal standard. Synthetic standards and international rock reference materials (G-2, JG-2, JR-2, JA-1, SY-3, AGV-2, BCR-2, and BHVO-2) were used for calibration. Analytical precision is better than 3% for major elements and better than 2% for trace and rare earth elements. Whole-rock geochemical data for the granitoids are summarized in Tables 1-3.

Zircon separation, mounting, and U–Pb dating were carried out at CSIR-NGRI, Hyderabad. Zircons were separated from 4 samples using standard mineral-separation procedures involving crushing, sieving, Wilfley table concentration, magnetic separation, and heavy-liquid separation. Handpicked zircons were mounted in epoxy, polished, and carbon-coated. Cathodoluminescence (CL) images were acquired using a TESCAN VEGA3 scanning electron microscope to guide spot selection. Zircon U–Pb analyses were performed using a 193 nm ArF excimer laser ablation system coupled with an Agilent 7800 Q-ICP-MS. Analysis is carried out with 30 µm spot size, with laser energy of 6.5 mJ while maintaining attenuation rate 25%, repetition rate of 10 Hz and ThO⁺/Th <1.

Zircon U–Pb data reduction and age calculation

Raw LA-ICP-MS data were processed using Iolite software, and common Pb correction was applied during data reduction. Zircon U–Pb ages were calculated from the ²⁰⁷Pb/²⁰⁶Pb, ²⁰⁶Pb/²³⁸U, and ²⁰⁷Pb/²³⁵U ratios. Zircon standards including TEMORA, GJ-1, and MT were analyzed during the analytical session, with GJ-1 used for calibration. U–Pb data evaluation, error propagation and age calculations were performed using the open source software package IsoplotR (Vermeesch, 2018). Final ages are reported as upper-intercept ages obtained by linear regression of the discordant arrays on the concordia diagram; within each array, analyses that are 95–100% concordant are termed concordant and anchor the upper-intercept regression. All analytical uncertainties are reported at the 1σ level.

RESULTS

Petrography

The study area comprises supracrustal rocks and granitoids that occur as discontinuous exposures aligned broadly along the regional structural grain. Among the supracrustal units, BIF is exposed as small hematitic mounds with characteristic cherry-red weathered surfaces, whereas amphibolite occurs as dark, medium- to coarse-grained bodies that are locally foliated. In thin section, the BIF shows alternating quartz-rich and magnetite-rich bands, while the amphibolite is composed mainly of hornblende, plagioclase, and biotite (Figure 3a, b). Other supracrustal lithologies, such as schist, ultramafic rock, and carbonatite, are present in the area but are not illustrated here.

The granitoids are represented in this study by granitic gneiss, granodiorite gneiss, and leucogranite. Granitic gneiss occurs as mesocratic, light grey to pink, coarse-grained, foliated rocks, with quartz, alkali feldspar, plagioclase, and biotite as the main minerals and zircon and apatite as accessories. The foliation is a megascopic gneissic fabric developed at outcrop and hand-specimen scale; owing to the coarse grain size and alkali-feldspar-dominated mineralogy, it is not resolved at thin-section scale. In thin section, alkali feldspar is commonly perthitic and locally shows albite stringers, flame perthite, and myrmekitic intergrowths (Figure 3c). Granodiorite gneiss is a foliated, medium-grained granitoid of variable grey shades, richer in plagioclase than alkali feldspar and composed of quartz, biotite, and hornblende. In migmatitic outcrops, the granodiorite gneiss appears as the mesosome, occurring between leucocratic segregations. Microscopic observations of granodiorite gneiss commonly show plagioclase with polysynthetic twinning and local sericitization, whereas

microcline exhibits cross-hatched twinning and perthitic textures (Figure 3d).

Leucogranite occurs as milky white to pale pink, medium- to coarse-grained small lensoid bodies dominated by quartz and perthitic alkali feldspar with minor biotite. Garnet-bearing

and magnetite-bearing variants are recognized, with garnet occurring as dark pink grains in some samples and magnetite as conspicuous black crystals in hand specimen. In thin section, the rocks are quartz–feldspar rich and contain large, fractured garnet megacrysts (Figure 3e). They locally show minor recrystallization of quartz and alteration of feldspar.

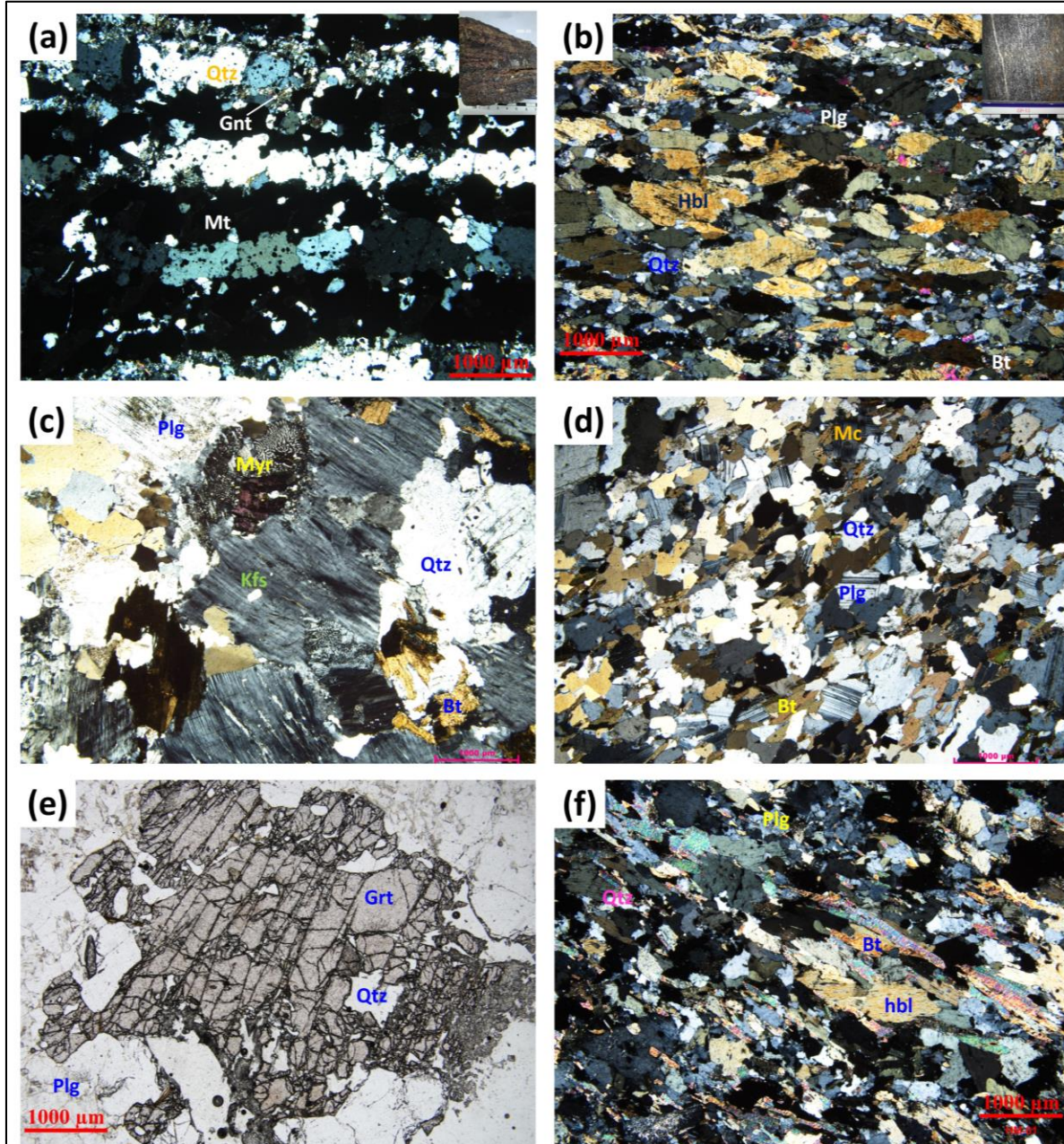


Figure 3. Microscopic images of Ghanapur belt rocks. (a) BIF in XPL showing alternate layers of quartz and magnetite; hand specimen in subset. (b) Foliated amphibolite in XPL with a subset showing hand specimen. (c) granitic gneiss in XPL showing perthitic K-feldspars and myrmekite texture. (d) Granodiorite grey gneiss in XPL showing lamellar twinning in plagioclase. (e) garnet-bearing leucogranite in PPL. (f) Mafic enclave in XPL with alignment of mica and amphiboles along the gneissic foliation trend. All images are under 2x magnification. Qtz- quartz, Plg- plagioclase, Mc- microcline, Kfs- Kfeldspar, Hbl- hornblende, Bt- biotite, Mt- magnetite, Grt- garnet, Gnt- grunerite and Myr- myrmekite

Mafic enclaves occur as mesocratic, medium-grained bodies enclosed within the granitoids and commonly show diffuse to gradational contacts with the host rock, locally with marginal reaction rims. Microscopically, the enclaves preserve a foliated texture defined by aligned platy mafic grains and elongated felsic minerals, superimposed on a hypidiomorphic-granular groundmass, and are composed of quartz, alkali feldspar, plagioclase, biotite, and hornblende (Figure 3f).

Geochemistry of granitoids

The granitoid suite is compositionally heterogeneous and includes granitic gneiss, granodiorite gneiss and leucogranite, spanning intermediate to highly evolved felsic compositions. The granitic gneisses are felsic, with SiO_2

contents of 67.0–78.0 wt%, and plot mainly in the granitic field, except for one sample that extends into the quartz monzonite field (Figure 4a). Major-element trends show decreasing $\text{Fe}_2\text{O}_3(\text{T})$, MgO , CaO , and TiO_2 with increasing SiO_2 , whereas K_2O increases with increasing SiO_2 . These rocks are metaluminous to slightly peraluminous, with A/CNK values of 0.88–1.09, and show high-K calc-alkaline to shoshonitic affinity, with magnesian character on the Fe -silica plot (Figure 4b–d). Chondrite-normalized (McDonough and Sun, 1995) REE patterns show variable fractionation with weakly negative to strongly positive Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.31\text{--}12.51$). Primitive mantle-normalized (McDonough and Sun, 1995) patterns show enrichment in large ion lithophile elements (LILEs) and depletion in Nb, Ta, and Ti (Figure 5a, b).

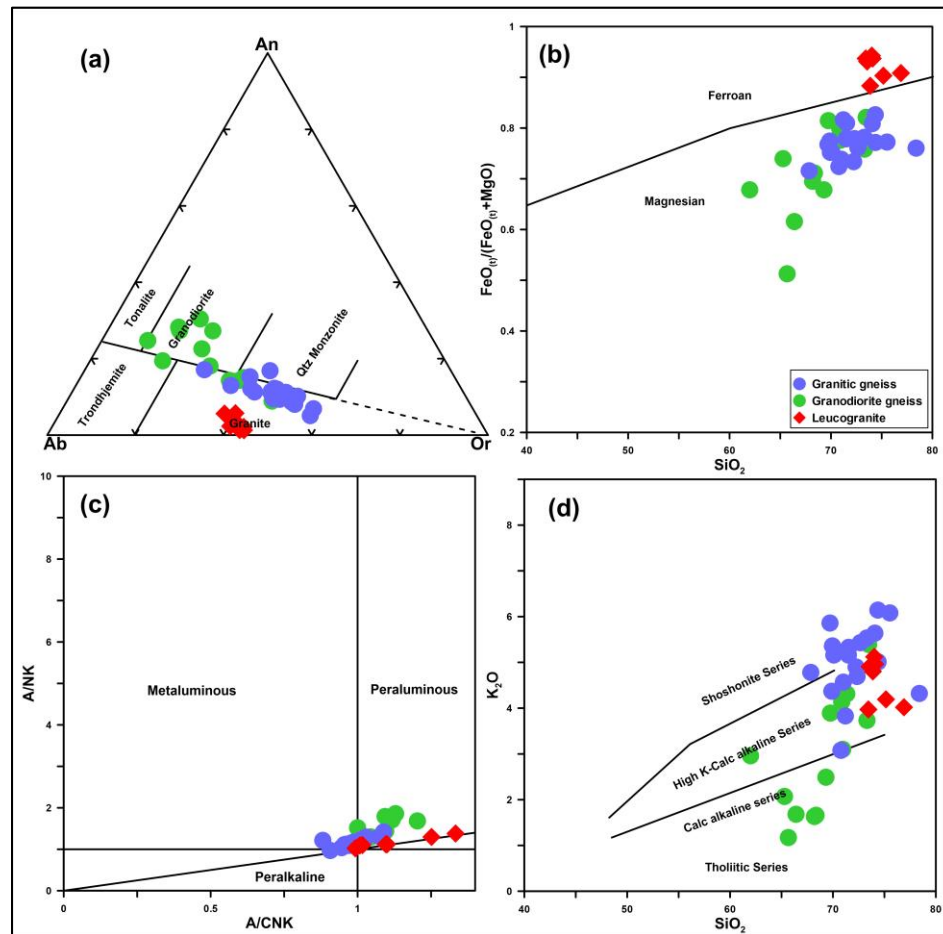


Figure 4. (a) Normative feldspar (Ab-An-Or) triangular diagram after O'Connor (1965) for classification of Ghanapur granitoids. (b) SiO_2 versus $\text{FeO}/(\text{FeO}+\text{MgO})$ granite tectonic discrimination diagram after Frost et al. (2001). (c) A/NK versus A/CNK diagram of Shand (1943) discriminating metaluminous, peraluminous and peralkaline compositions. (d) SiO_2 versus K_2O after Peccerillo and Taylor (1976). Symbols for the three granitoid suites are given in the legend and same code is maintained for all geochemical discrimination.

Table 1. Major oxide concentrations, trace element, REE concentrations and ratios of granitic gneisses from Ghanapur.

Granitic gneisses	GP-06	GP-11	GP-15	HM-02	KD-02	ND-02	PN-07	PN-08	SNR-01	SNR-02	SR-01	SR-03	TG-04	ZG-03	ZG-04	ZG-05	ZG-12	ZG-13	ZG-16
<i>Major oxides (wt%)</i>																			
SiO ₂	71.21	69.95	69.88	73.28	71.54	74.07	74.37	69.71	67.85	78.40	74.38	75.53	72.34	70.78	72.24	71.01	71.50	72.65	70.08
TiO ₂	0.22	0.31	0.25	0.17	0.19	0.06	0.04	0.18	0.33	0.03	0.15	0.11	0.41	0.31	0.18	0.30	0.23	0.14	0.41
Al ₂ O ₃	15.05	14.76	15.53	14.11	15.29	14.31	13.93	15.04	14.66	12.07	13.90	13.08	13.31	15.64	15.27	15.30	14.96	14.90	15.15
Fe ₂ O ₃ (T)	2.32	3.37	2.90	1.95	1.80	1.22	0.95	3.12	3.69	0.60	1.39	1.17	3.53	3.06	1.99	3.58	2.23	1.31	2.74
MnO	0.03	0.03	0.02	0.01	0.03	0.01	0.01	0.03	0.07	0.01	0.02	0.01	0.03	0.03	0.01	0.04	0.02	0.02	0.04
MgO	0.47	1.00	0.76	0.49	0.38	0.26	0.18	0.85	1.32	0.17	0.37	0.31	0.90	1.05	0.65	1.14	0.57	0.37	0.81
CaO	1.82	1.71	2.27	1.18	1.76	1.16	0.95	1.65	3.39	1.32	1.28	0.70	1.44	2.46	1.75	1.96	1.76	1.46	1.74
Na ₂ O	3.95	3.10	3.69	3.04	3.38	3.01	2.72	3.02	3.25	2.91	3.31	2.81	3.10	4.43	3.92	3.80	3.36	3.36	3.48
K ₂ O	3.83	5.36	4.37	5.53	5.33	5.64	6.14	5.86	4.78	4.32	5.01	6.08	4.69	3.08	4.90	4.57	5.16	5.43	5.16
P ₂ O ₅	0.10	0.18	0.15	0.05	0.08	0.09	0.05	0.21	0.27	0.01	0.04	0.04	0.07	0.11	0.12	0.24	0.08	0.15	0.18
Total	99.00	99.77	99.82	99.81	99.78	99.83	99.34	99.67	99.61	99.84	99.85	99.84	99.82	100.95	101.03	101.94	99.87	99.79	99.79
Fe#	71.33	62.94	65.79	66.73	70.48	70.28	72.68	64.91	58.49	64.01	65.44	65.54	66.41	59.49	60.68	61.28	66.35	64.09	63.03
Mg#	28.67	37.06	34.21	33.27	29.52	29.72	27.32	35.09	41.51	35.99	34.56	34.46	33.59	40.51	39.32	38.72	33.65	35.91	36.97
A/NK	1.30	1.16	1.29	1.09	1.17	1.10	1.04	1.12	1.22	1.11	1.11	0.97	1.14	1.42	1.16	1.22	1.17	1.13	1.17
A/KCN	1.07	0.98	1.03	0.97	0.99	0.98	0.95	0.96	0.88	0.96	0.98	0.91	0.98	1.09	0.98	1.01	0.99	0.98	0.99
K ₂ O/Na ₂ O	0.97	1.73	1.18	1.82	1.58	1.87	2.26	1.94	1.47	1.48	1.51	2.16	1.51	0.70	1.25	1.20	1.54	1.62	1.48
<i>TE (ppm)</i>																			
Cs	1.08	1.73	1.21	1.89	0.53	1.07	1.30	3.47	1.05	0.76	1.29	0.63	1.45	1.29	1.21	1.90	1.64	0.68	2.02
Rb	109.16	210.62	160.85	215.63	181.97	179.46	199.17	280.69	138.06	130.75	187.33	203.06	218.91	145.88	168.03	193.79	177.37	161.36	194.68
Ba	1869.65	748.99	773.05	585.13	1379.77	847.67	970.21	2492.59	2491.59	903.28	693.62	1014.59	707.53	631.22	747.96	865.55	1141.18	1135.04	759.87
Nb	5.77	8.80	12.65	13.54	7.88	5.93	4.90	8.17	8.45	5.12	15.87	14.53	37.17	11.25	11.77	17.29	7.06	4.41	7.18
Sr	418.72	260.03	270.68	143.67	344.41	231.47	286.24	402.98	530.99	249.38	86.95	70.67	96.48	345.89	287.63	286.85	246.46	298.71	224.60
Zr	138.34	273.54	147.16	253.60	124.83	117.01	174.34	49.55	157.35	85.78	120.90	147.75	240.36	145.63	115.06	160.84	90.74	132.02	342.30
Y	10.40	43.28	20.28	41.25	6.48	21.90	31.52	9.19	35.91	2.04	8.38	27.85	7.66	18.01	19.73	30.47	4.83	14.04	31.27
La	37.78	62.29	23.02	82.47	14.57	15.21	73.33	12.01	36.86	15.55	62.59	44.66	31.36	49.17	30.72	34.21	12.54	63.41	69.88
Ce	63.25	119.83	39.99	155.43	24.68	27.80	131.89	18.87	67.04	23.65	117.90	84.73	55.02	86.80	52.89	59.04	20.56	115.73	125.13
Pr	6.19	12.70	4.13	16.27	2.68	2.91	14.65	1.89	8.87	2.06	12.39	9.08	5.60	9.51	5.93	6.65	2.08	11.79	12.61

Continued

Nd	20.84	43.06	14.31	54.70	9.00	8.90	43.74	5.86	32.46	5.57	42.04	31.88	19.11	31.10	19.08	21.87	7.82	39.44	41.88
Sm	3.37	8.56	2.89	10.83	1.86	1.94	7.60	1.27	6.46	0.77	6.56	6.42	3.47	5.89	3.82	4.22	1.61	6.06	7.25
Eu	2.47	1.34	1.36	1.16	3.06	1.84	2.34	4.76	5.33	2.01	1.48	1.65	1.35	0.99	0.80	0.86	1.55	1.89	1.38
Gd	4.12	9.74	3.16	11.53	1.30	1.54	6.23	1.00	4.29	0.79	7.00	7.36	3.82	5.00	3.57	4.10	1.69	7.37	8.69
Tb	0.38	1.25	0.42	1.35	0.23	0.37	1.03	0.20	0.88	0.09	0.53	0.86	0.35	0.63	0.58	0.67	0.18	0.69	0.90
Dy	1.72	7.03	2.70	6.84	1.24	2.92	5.54	1.41	5.65	0.37	1.69	4.43	1.40	3.37	3.47	4.34	0.83	2.97	4.84
Ho	0.29	1.20	0.53	1.13	0.27	0.73	1.12	0.31	1.31	0.08	0.24	0.77	0.22	0.62	0.67	0.97	0.13	0.42	0.85
Er	0.81	3.21	1.63	3.07	0.56	1.75	2.27	0.67	2.85	0.19	0.67	1.98	0.60	1.56	1.73	2.86	0.36	1.06	2.35
Tm	0.13	0.49	0.27	0.46	0.08	0.30	0.30	0.11	0.46	0.04	0.10	0.26	0.09	0.20	0.21	0.25	0.05	0.18	0.34
Yb	0.76	2.85	1.65	2.69	0.43	1.68	1.62	0.60	2.50	0.22	0.55	1.47	0.65	1.23	1.23	0.91	0.28	1.41	1.73
Lu	0.12	0.43	0.24	0.43	0.06	0.24	0.25	0.09	0.36	0.04	0.10	0.24	0.12	0.16	0.16	0.11	0.05	0.22	0.26
Hf	3.59	7.82	4.26	7.66	3.34	3.73	5.96	1.34	4.32	2.78	3.75	4.74	6.72	3.76	3.26	4.15	2.13	3.76	9.04
Ta	0.31	2.71	1.06	1.40	0.53	1.05	1.11	1.06	0.49	0.18	0.44	0.49	0.83	1.24	1.38	1.55	0.27	0.66	0.77
Th	19.02	48.17	11.86	64.08	14.35	11.30	56.68	10.26	14.44	9.00	39.90	27.55	35.87	20.84	16.39	16.45	3.54	40.17	47.76
U	2.86	15.19	8.13	40.69	1.25	7.91	23.40	8.52	1.83	10.70	3.24	4.75	5.32	4.35	13.54	12.30	0.80	7.09	37.35
Pb	39.56	65.25	49.02	87.57	34.95	60.53	73.11	63.14	39.58	43.02	45.96	75.61	39.71	39.22	54.47	46.16	30.21	59.53	64.57
Sc	3.34	5.89	4.76	4.73	2.97	2.08	1.99	6.00	8.36	0.63	2.58	2.93	8.66	6.85	4.26	7.47	2.51	2.71	6.66
V	21.55	40.02	41.45	20.56	21.32	6.75	8.89	29.04	50.58	4.69	7.09	6.31	25.09	32.64	15.25	31.42	27.04	18.87	19.44
Co	4.53	7.52	5.20	4.11	4.19	2.60	1.89	7.15	9.42	1.38	1.72	1.70	5.86	8.39	3.74	8.67	5.17	3.14	5.66
Ni	12.34	17.24	10.66	10.44	9.81	5.86	6.05	21.56	35.45	5.53	2.17	1.99	4.57	244.46	84.84	180.04	11.24	7.06	14.65
Ga	16.07	18.00	17.74	16.66	16.26	13.35	14.61	18.48	18.73	12.74	14.92	12.83	15.69	20.78	15.65	19.55	15.41	15.46	17.94
Ba/Rb	17.13	3.56	4.81	2.71	7.58	4.72	4.87	8.88	18.05	6.91	3.70	5.00	3.23	4.33	4.45	4.47	6.43	7.03	3.90
Sr/Y	40.26	6.01	13.35	3.48	53.11	10.57	9.08	43.87	14.79	122.29	10.37	2.54	12.60	19.20	14.58	9.41	51.06	21.28	7.18
Nb/Ta	18.65	3.25	11.88	9.69	14.78	5.65	4.41	7.73	17.37	28.21	36.42	29.49	44.87	9.11	8.55	11.18	25.93	6.64	9.35
Lu/Hf	0.03	0.06	0.06	0.06	0.02	0.06	0.04	0.07	0.08	0.01	0.03	0.05	0.02	0.04	0.05	0.03	0.02	0.06	0.03
ΣREE	142.23	273.99	96.31	348.38	60.02	68.12	291.92	49.05	175.32	51.42	253.83	195.79	123.15	196.23	124.86	141.06	49.72	252.65	278.09
La _n /Yb _n	33.57	14.78	9.40	20.70	23.05	6.12	30.55	13.54	9.97	48.31	76.83	20.51	32.76	27.09	16.88	25.33	30.43	30.31	27.32
Eu/Eu*	2.02	0.45	1.37	0.32	5.72	3.15	1.01	12.51	2.92	7.84	0.66	0.73	1.13	0.54	0.65	0.62	2.85	0.86	0.53
Rb/Sr	0.26	0.81	0.59	1.50	0.53	0.78	0.70	0.70	0.26	0.52	2.15	2.87	2.27	0.42	0.58	0.68	0.72	0.54	0.87

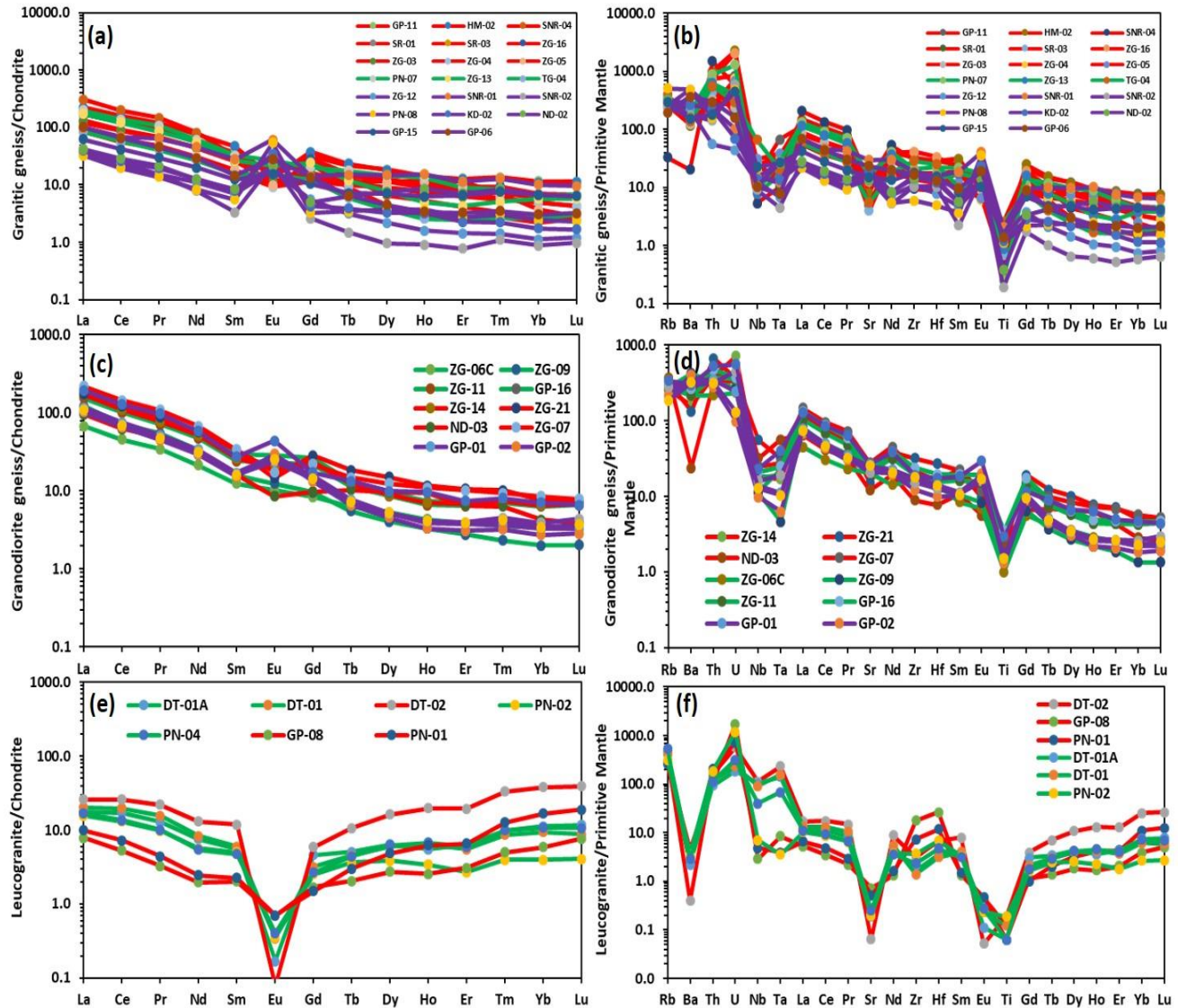


Figure 5. Normalized REE and trace element patterns for granitoids of Ghanapur. (a) Chondrite normalized REE patterns of granitic gneiss. (b) Primitive mantle normalized trace element spider plot for granitic gneiss. (c) Chondrite normalized REE patterns of granodiorite gneiss. (d) Primitive mantle normalized multi-element spider diagram granodiorite gneiss. (e) Chondrite normalized REE patterns of Leucogranite. (f) Primitive mantle normalized multi-element spider diagram Leucogranite. Normalization values are used after McDonough and Sun (1995)

The granodiorite gneisses are silica-rich to moderately silica-rich, with SiO_2 ranging from 61.98 to 73.31 wt%, and mostly plot in the granodioritic field, with a few samples extending into the granitic field (Figure 4a). These rocks show calc-alkaline to high-K calc-alkaline affinities and are generally weakly peraluminous, with A/CNK values of 0.96–1.20 (Figure 4c, d). Their trace-element patterns show moderate REE abundances and variable Eu behavior ($\text{Eu}/\text{Eu}^* = 0.47\text{--}2.07$), and show enrichment in LILE and depletion in high field strength elements (HFSE) elements. Relative to the granitic gneisses, the granodiorite gneisses record a stronger role for plagioclase in their petrogenesis, expressed as a wider spread in Eu anomalies, negative values reflecting

residual or fractionating plagioclase and positive values reflecting plagioclase accumulation (Figure 5c, d).

The leucogranites are the most evolved rocks in the assemblage. They are highly silica-rich, with SiO_2 values of 73.44–76.90 wt %, and all samples plot in the granitic field (Figure 4a). These rocks are peraluminous to strongly peraluminous, with A/CNK values of 0.99–1.33, and show high-K calc-alkaline to potassic affinities, consistent with highly evolved felsic melts (Figure 4c, d). Their trace-element patterns are marked by low REE contents and strong negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.01\text{--}0.38$), with enrichment in Rb, Th, U, Zr, and Hf, –

Table 2. Major oxide concentrations, trace element and REE concentrations and ratios of granodiorite gneisses from Ghanapur.

Granodiorite gneiss	ZG-06C	GP-01	GP-02	GP-04	GP-07	GP-16	KD-03	ND-03	ZG-07	ZG-09	ZG-11	ZG-14	ZG-21
<i>Major oxides (wt%)</i>													
SiO ₂	73.47	70.85	71.34	69.31	70.94	65.28	65.67	66.40	61.98	73.31	68.19	68.36	69.74
TiO ₂	0.16	0.24	0.21	0.25	0.24	0.57	0.47	0.40	0.53	0.22	0.46	0.45	0.33
Al ₂ O ₃	14.67	15.21	15.27	15.18	15.43	15.70	15.61	15.19	16.95	14.37	14.78	14.59	14.58
Fe ₂ O ₃ (T)	2.14	2.65	2.07	3.70	2.54	6.15	4.09	5.55	5.74	2.58	4.61	5.04	4.49
MnO	0.02	0.03	0.03	0.07	0.04	0.08	0.08	0.05	0.10	0.03	0.06	0.07	0.04
MgO	0.42	0.60	0.52	1.58	0.66	1.95	3.50	3.12	2.45	0.74	1.82	1.84	0.92
CaO	1.36	2.12	2.11	3.49	2.48	4.04	3.83	2.66	4.28	1.97	3.90	3.57	2.01
Na ₂ O	3.60	3.74	3.71	3.55	4.19	3.71	5.07	4.56	4.66	3.93	4.06	3.98	3.66
K ₂ O	5.39	4.14	4.31	2.49	3.10	2.07	1.17	1.68	2.96	3.73	1.63	1.66	3.89
P ₂ O ₅	0.10	0.09	0.07	0.08	0.07	0.15	0.22	0.23	0.19	0.08	0.22	0.21	0.14
Total	101.33	99.67	99.64	99.70	99.69	99.70	99.71	99.84	99.84	100.96	99.73	99.77	99.80
Fe#	71.97	69.00	66.74	54.13	65.98	61.38	37.07	47.27	54.14	63.73	56.07	57.99	71.10
Mg#	28.03	31.00	33.26	45.87	34.02	38.62	62.93	52.73	45.86	36.27	43.93	42.01	28.90
A/NK	1.09	1.30	1.28	1.71	1.44	1.86	1.75	1.68	1.52	1.26	1.79	1.78	1.30
A/KCN	0.96	1.04	1.03	1.12	1.10	1.13	1.11	1.20	1.00	1.02	1.09	1.12	1.05
K ₂ O/Na ₂ O	1.50	1.11	1.16	0.70	0.74	0.56	0.23	0.37	0.64	0.95	0.40	0.42	1.06
<i>TE (ppm)</i>													
Cs	0.88	1.96	1.32	3.50	1.55	2.56	6.07	5.31	2.90	1.28	2.53	2.07	1.76
Rb	190.01	115.41	118.65	154.20	102.38	171.56	185.09	202.02	158.23	152.31	146.70	139.34	175.68
Ba	1091.48	1751.87	2027.32	1473.04	1619.85	1369.49	1528.52	120.78	1117.83	2161.32	1109.87	807.90	683.87
Nb	6.51	7.63	5.51	8.89	7.15	11.59	12.94	17.81	13.89	6.21	9.90	10.42	31.20
Sr	361.82	442.24	432.08	367.41	451.66	402.26	481.82	216.05	496.35	473.24	382.34	327.71	270.27
Zr	135.83	144.13	101.71	127.41	147.50	198.20	159.44	73.46	176.04	134.32	159.47	157.56	263.82
Y	10.54	11.33	10.04	13.58	13.33	23.53	21.17	22.11	27.11	7.87	19.98	21.33	35.27
La	24.87	43.01	37.82	39.64	40.28	73.48	71.58	35.62	81.15	45.18	56.94	62.29	63.80
Ce	43.90	72.35	62.94	66.87	67.06	124.13	123.36	60.01	137.91	69.97	97.88	105.67	117.07
Pr	4.75	7.14	6.18	6.54	6.57	12.32	13.37	6.77	14.86	7.45	9.82	10.72	12.16
Nd	15.28	24.06	20.82	21.95	22.25	42.00	41.60	20.78	48.18	23.85	33.77	36.62	41.15
Sm	2.90	3.92	3.55	3.74	3.70	6.76	6.31	3.75	7.84	3.62	5.58	6.04	7.71
Eu	0.90	2.42	2.59	2.07	2.19	2.51	3.83	0.74	1.53	1.09	2.01	1.72	1.27
Gd	2.59	4.78	4.10	4.49	4.40	8.02	5.29	2.99	6.76	2.95	6.63	7.26	8.73
Tb	0.36	0.45	0.39	0.41	0.42	0.79	0.77	0.60	0.86	0.32	0.66	0.72	1.07
Dy	2.02	1.98	1.74	1.96	2.01	3.69	3.78	3.70	4.75	1.55	3.28	3.41	5.77
Ho	0.37	0.33	0.28	0.35	0.35	0.63	0.82	0.80	0.94	0.28	0.56	0.60	0.99
Er	0.96	0.94	0.77	0.99	0.97	1.78	1.86	1.62	2.55	0.69	1.61	1.70	2.64
Tm	0.13	0.13	0.12	0.16	0.15	0.27	0.29	0.23	0.34	0.08	0.25	0.27	0.37
Yb	0.85	0.81	0.68	0.97	0.86	1.71	1.76	1.07	2.13	0.50	1.58	1.61	1.89
Lu	0.12	0.13	0.11	0.17	0.14	0.27	0.25	0.15	0.30	0.08	0.25	0.26	0.28
Hf	3.94	3.82	2.70	3.45	3.78	5.26	4.69	2.10	4.43	3.43	4.19	4.11	7.32
Ta	0.91	0.45	0.25	0.76	0.41	0.99	1.62	2.26	1.11	0.18	0.76	0.68	1.28
Th	13.89	22.25	18.12	20.55	20.14	30.30	33.55	20.19	28.72	20.79	23.88	24.87	42.81
U	4.26	4.35	1.76	8.14	2.35	6.70	10.07	5.99	6.09	4.81	6.32	12.86	6.31
Pb	57.64	42.97	42.52	30.57	37.60	21.10	33.41	22.34	32.05	40.42	23.05	26.98	62.73
Sc	3.41	3.67	2.97	3.32	3.47	8.70	8.41	9.73	10.24	3.58	6.94	7.35	7.26
V	8.82	27.72	20.79	27.36	23.90	65.96	67.29	54.04	81.19	25.06	57.69	60.03	52.13
Co	4.33	5.20	3.78	6.10	4.81	12.87	12.31	10.01	16.87	5.30	10.39	11.16	7.25
Ni	109.22	15.22	10.01	12.19	12.40	44.25	32.08	20.24	466.16	178.35	29.71	29.78	17.33
Ga	17.26	17.33	16.36	22.01	16.78	19.34	19.61	24.12	24.44	17.99	16.70	17.42	21.33
Ba/Rb	5.74	15.18	17.09	9.55	15.82	7.98	8.26	0.60	7.06	14.19	7.57	5.80	3.89
Sr/Y	34.32	39.02	43.05	27.05	33.88	17.10	22.76	9.77	18.31	60.15	19.14	15.37	7.66
Nb/Ta	7.19	17.15	21.81	11.68	17.35	11.67	7.99	7.87	12.47	33.64	12.97	15.42	24.44
Lu/Hf	0.03	0.03	0.04	0.05	0.04	0.05	0.05	0.07	0.07	0.02	0.06	0.06	0.04
ΣREE	99.99	162.44	142.09	150.32	151.35	278.34	274.87	138.82	310.10	157.61	220.83	238.90	264.89
La _n /Yb _n	19.87	35.91	37.76	27.71	31.73	29.01	27.56	22.57	25.71	61.38	24.36	26.09	22.80
Eu/Eu*	0.99	1.70	2.07	1.55	1.66	1.04	1.97	0.65	0.63	0.99	1.01	0.79	0.47
Rb/Sr	0.53	0.26	0.27	0.42	0.23	0.43	0.38	0.94	0.32	0.32	0.38	0.43	0.65

Table 3. Major oxide concentrations, trace element and REE concentrations and ratios of leucogranites from Ghanapur.

	<i>Leucogranite (garnetiferous)</i>				<i>Leucogranite (magnetite bearing)</i>		
	DT-01	DT-01A	PN-04	DT-02	GP-08	PN-01	PN-02
<i>Major oxides (wt%)</i>							
SiO ₂	73.57	74.10	75.16	73.44	73.87	76.90	74.02
TiO ₂	0.02	0.01	0.01	0.03	0.01	0.02	0.03
Al ₂ O ₃	15.58	15.52	15.42	16.19	15.35	13.43	14.66
Fe ₂ O ₃ (T)	0.60	0.66	0.83	1.16	0.42	0.44	0.91
MnO	0.07	0.08	0.11	0.18	0.00	0.03	0.02
MgO	0.04	0.04	0.08	0.07	0.05	0.04	0.05
CaO	0.21	0.21	0.32	0.29	0.85	0.77	0.35
Na ₂ O	4.44	4.31	3.79	3.94	4.58	4.27	4.47
K ₂ O	4.91	4.96	4.19	3.97	4.80	4.02	5.12
P ₂ O ₅	0.02	0.03	0.03	0.01	0.02	0.04	0.04
Total	99.46	99.92	99.94	99.28	99.95	99.96	99.67
Fe#	88.32	89.27	83.95	89.31	80.89	84.72	90.17
Mg#	11.68	10.73	16.05	10.69	19.11	15.28	9.83
A/NK	1.12	1.12	1.30	1.38	1.10	1.09	1.03
A/KCN	1.10	1.10	1.25	1.33	1.02	1.01	0.99
K ₂ O/Na ₂ O	1.11	1.15	1.11	1.01	1.05	0.94	1.15
<i>TE (ppm)</i>							
Cs	0.88	0.57	1.81	0.46	0.96	0.98	3.20
Rb	249.04	238.16	292.75	222.35	158.10	140.52	170.80
Ba	15.70	11.36	14.75	2.04	23.26	12.06	18.60
Nb	51.85	58.00	22.56	63.10	1.60	2.72	3.92
Sr	4.19	3.42	4.52	1.14	12.94	9.15	3.47
Zr	11.68	14.09	19.51	25.62	149.04	61.51	30.57
Y	20.51	22.80	20.93	52.81	8.27	19.30	10.50
La	7.43	6.53	6.18	9.50	2.91	3.65	5.77
Ce	18.86	16.44	13.13	25.00	5.04	6.91	12.49
Pr	2.13	1.76	1.39	3.03	0.45	0.60	1.35
Nd	5.92	5.39	3.89	9.35	1.39	1.73	4.00
Sm	1.38	1.30	1.10	2.74	0.47	0.52	1.15
Eu	0.03	0.01	0.04	0.01	0.06	0.06	0.03
Gd	0.97	1.41	0.82	1.81	0.52	0.46	0.77
Tb	0.26	0.30	0.26	0.62	0.12	0.18	0.21
Dy	2.33	2.45	2.39	6.22	1.04	1.85	1.47
Ho	0.54	0.47	0.58	1.69	0.22	0.53	0.29
Er	1.38	1.59	1.50	4.82	0.77	1.64	0.67
Tm	0.31	0.35	0.36	1.20	0.18	0.46	0.14
Yb	2.34	2.79	2.69	9.52	1.47	4.16	0.99
Lu	0.34	0.45	0.40	1.50	0.29	0.72	0.16
Hf	0.84	0.95	1.27	1.84	7.15	3.17	1.85
Ta	6.27	5.67	2.73	9.24	0.34	0.16	0.14
Th	7.55	5.94	7.25	10.12	6.70	13.27	11.82
U	4.19	3.33	5.72	10.36	31.29	13.17	21.45
Pb	19.35	24.25	25.97	26.57	62.29	48.36	20.51
Sc	4.86	5.29	14.73	14.53	1.31	2.06	1.68
V	1.33	1.22	1.34	1.26	1.36	1.76	1.51
Co	0.61	0.62	0.80	0.59	0.61	0.79	0.62
Ni	2.72	2.01	3.20	2.64	1.89	3.42	2.88
Ga	26.28	24.78	25.76	31.11	14.82	16.57	24.32
Ba/Rb	0.06	0.05	0.05	0.01	0.15	0.09	0.11
Sr/Y	0.20	0.15	0.22	0.02	1.57	0.47	0.33
Nb/Ta	8.27	10.23	8.27	6.83	4.74	17.35	27.33
Lu/Hf	0.40	0.47	0.32	0.81	0.04	0.23	0.08
ΣREE	44.22	41.25	34.71	77.00	14.93	23.47	29.50
La _N /Yb _N	2.15	1.58	1.55	0.67	1.33	0.59	3.96
Eu/Eu*	0.08	0.03	0.11	0.01	0.37	0.38	0.09
Rb/Sr	59.42	69.58	64.72	194.44	12.21	15.35	49.16

coupled with depletion in Ba, Sr, Eu, and Ti. Their low Mg# values (9.83–19.11), and very low CaO contents are consistent with highly evolved felsic compositions. Garnet-bearing varieties show flatter HREE patterns and stronger Eu depletion, whereas non-garnetiferous samples retain relatively higher HREE concentrations (Figure 5e, f).

Zircon geochronology of granitoids

Granitic gneiss

Many zircons show variable metamictization; therefore, upper-intercept concordia ages are preferred for age interpretation. Sample TG-04 (17.77884, 79.42641) is collected from an inactive granite quarry at Theegaram village along the western margin of the Ghanapur belt. The sample is coarse-grained, and well-foliated. Separated zircons are mainly elongated to prismatic, 80–100 μm long, with few slightly rounded grains having 50–80 μm length. CL images show disrupted concentric oscillatory zoning and irregular homogeneous domains, indicating modification of the original magmatic zoning (Figure 6a). Seventeen analyses record U contents of 7–896 ppm, Th contents of 5–335 ppm, and Th/U ratios of 0.10–1.05. The analyses define a discordant array; regression of the array yields an upper-

intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2697 ± 24 Ma (MSWD = 6.0), anchored by four 95–100% concordant spots (spot locations shown in Fig. 6a) and taken as the best possible estimate of the crystallization age (Figure 7a). The elevated MSWD is attributed to minor scatter from variable, metamictization-related Pb-loss; ablation spots were sited on preserved oscillatory-zoned magmatic domains and away from recrystallized or resorbed zones.

Sample SR-01 (17.77354, 79.43184) is collected from a granitoid mound near Suraram village in the western part of the Ghanapur belt. The sample is coarse-grained, foliated, and banded. Separated zircons are mainly tabular, with a few elongated grains, and show rounded to prismatic terminations (Figure 6b). CL images show zoning-free grains with dark homogeneous interiors and minor bright patches, whereas a few grains preserve resorbed and distorted zoning (Figure 6b). Twenty-nine analyses yield U contents of 27–321 ppm, Th contents of 7–83 ppm, and Th/U ratios of 0.11–0.82. The analyses define a discordant array; regression of the array yields an upper-intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2607 ± 41 Ma (MSWD = 2.6), anchored by two 95–100% concordant spots, interpreted as the crystallization age (Figure 7b).

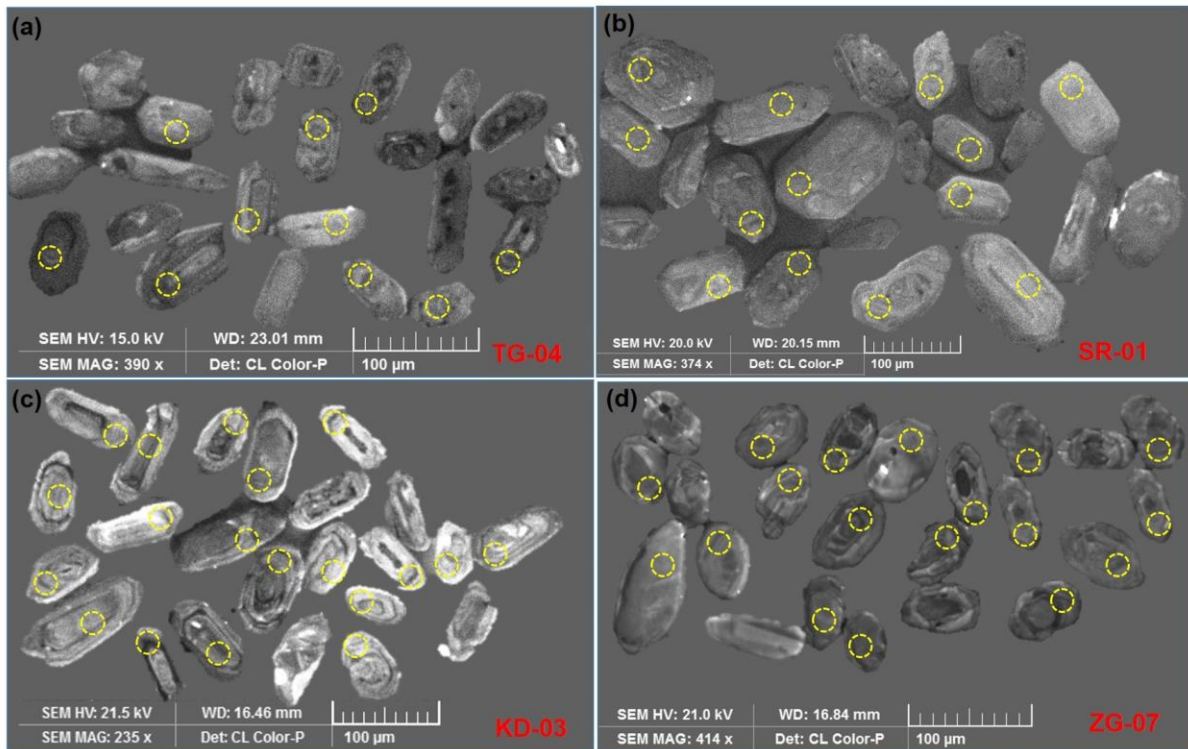


Figure 6. CL-SEM images of representative zircons from granitoids of Ghanapur (a-b) Zircons from granitic gneiss samples. (c-d) zircons from granodiorite gneiss samples. Yellow circles represent ablation spots.

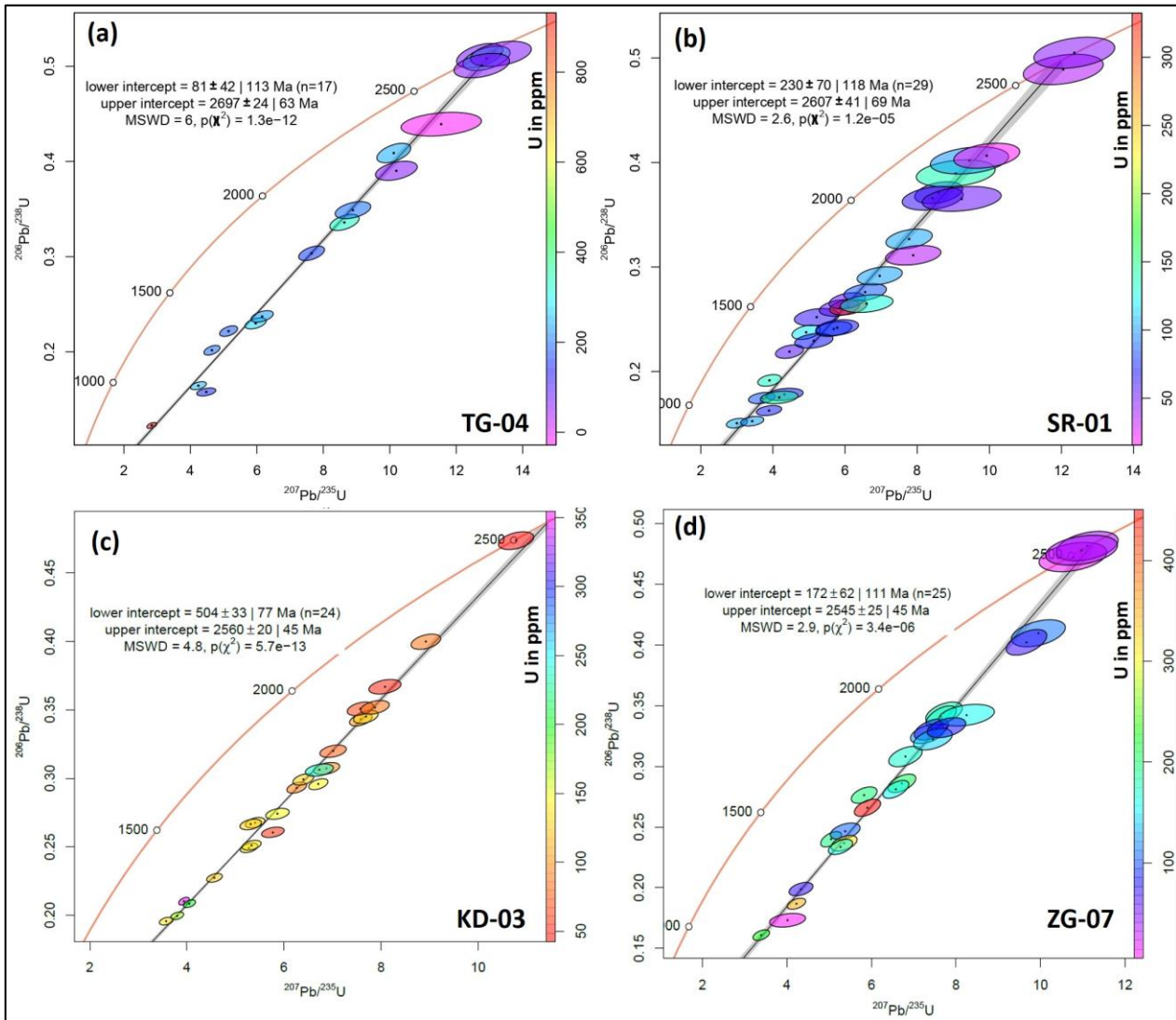


Figure 7. U-Pb concordia diagrams for zircons analysed. **(a-b)** Granitic gneiss samples. **(c-d)** Granodiorite gneiss samples.

Granodiorite gneiss

Two granodiorite gneiss samples appearing as mesosome in migmatitic outcrops were dated. Sample KD-03 (17.70269, 79.55881) is melanocratic gneiss collected from Kondur Gutta granite hill in the southern part of the Ghanapur belt. Separated zircons are mostly elongated to subrounded, 80–120 μm long, and commonly preserve rounded pyramidal terminations (Figure 6c). CL images show resorption of primary magmatic zoning, and some grains preserve dark xenocrystic cores mantled by thin dark rims, together with dark homogeneous domains, indicating post-magmatic recrystallization in the presence of aqueous fluids (Figure 6c). Twenty-four analyses record U contents of 54–343 ppm, Th contents of 33–190 ppm, and Th/U ratios of 0.21–1.11. The analyses define a discordant array; regression yields an upper-intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2560 ± 20 Ma (MSWD =

4.8), anchored by a single 99% concordant spot and interpreted as the crystallization age (Figure 7c).

Sample ZG-07 (17.771386, 79.481862) collected from the temple hill of Zaffergadh, is a mesocratic quartz-monzodiorite gneiss representing the mesosome portion of the migmatite outcrop. Separated zircons are mostly rounded to subrounded, 50–100 μm long, and pale yellow to light brown in color. CL images show resorption of primary zoning, with non-luminescent to weakly luminescent homogeneous xenocrystic cores and deformed rim zoning, indicating recrystallization of igneous zircon in the presence of magmatically derived aqueous fluids (Figure 6d). Twenty-five analyses yield U contents of 22–434 ppm, Th contents of 4–465 ppm, and Th/U ratios of 0.19–1.10. The analyses define a discordant array; regression of the array yields an upper-intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2545 ± 25 Ma (MSWD =

2.9), anchored by three 95–100% concordant spots, interpreted as the crystallization age (Figure 7d).

DISCUSSION

Petrogenesis and source characteristics of the granitoid suite

The granitoid suite of the Ghanapur area records progressive crustal reworking and magmatic differentiation rather than derivation from a single homogeneous source. Its petrographic variability, together with the compositional transition from granitic gneiss to granodiorite gneiss and leucogranite, indicates that the suite formed through multiple melt-generation and reworking events within a heterogeneous Archean crust. Major-element trends, including decreasing Fe_2O_3 (T), MgO , CaO , and TiO_2 with increasing SiO_2 , point to progressive fractionation during magma evolution, whereas enrichment in LILEs together with Nb–Ta–Ti depletion reflects strong influence of the source (Figure 5). These signatures are consistent with evolved Archean granitoid systems formed through prolonged crustal melting and reworking (Moyen, 2011; Laurent et al., 2014).

The granitic gneisses preserve the clearest evidence for mixed source contributions. In the Archean granitoid classification diagram, the granitic gneisses plot mainly in the biotite two-mica granite field, with two samples extending into the sanukitoid field, indicating a dominantly crustal granite lineage with local mafic imprint or hybridization (Figure 8a ; Moyen, 2011; Laurent et al., 2014). This is

further supported by their position in the Laurent et al. (2014) source-discrimination diagram, which suggests a contribution from metasedimentary protoliths alongside tonalitic and high-K mafic sources; in the absence of radiogenic-isotope data, this metasedimentary contribution is treated as suggestive (Figure 8b). The geochemical variability of the suite together with their weakly metaluminous to slightly peraluminous character and variable Eu anomaly, is consistent with melting of heterogeneous crustal protoliths at different crustal levels. Experimental studies have shown that peraluminous granites may form by dehydration melting of metasedimentary rocks, whereas more metaluminous, Ba–Sr–Ni–Cr-enriched compositions are commonly linked to partial melting of mafic lower crust or tonalitic material (Patiño Douce and Johnston, 1991; Chappell and White, 2001; Clemens, 2003).

In the Laurent et al. (2014) ternary classification diagram for Archean granitoids, the dated granodiorite gneissic samples plot in the sanukitoid field (Figure 8a), whereas the remaining samples fall in the overlapping zone between the sanukitoid, biotite two-mica granite, and TTG fields. The geochemical characteristics of the granodioritic gneisses, including their calc-alkaline to high-K calc-alkaline affinity, elevated Mg#, high Ba+Sr, and LILE and LREE enrichment, are consistent with a sanukitoid affinity and indicate derivation from an enriched mantle source wherein the mantle wedge was previously metasomatized by fluids and/or melts derived from a subducting slab, including oceanic crustal and sedimentary components (Laurent et al., 2014).

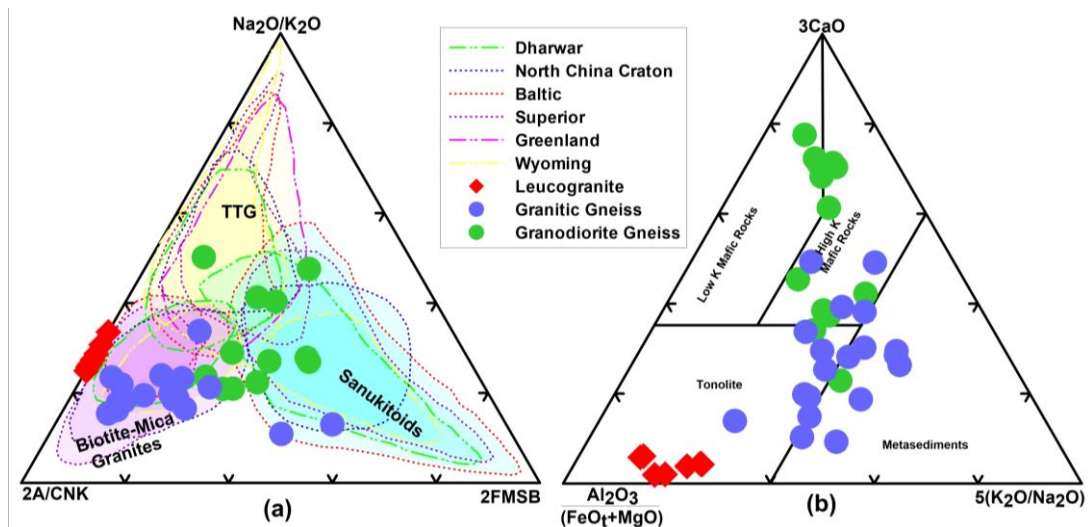


Figure 8. (a) Ternary synthetic classification diagram of Laurent et al. (2014), showing the petrogenetic processes involved in the genesis of the Late Archean granites. **(b)** Ternary source-discrimination diagram of the Ghanapur granitoids after Laurent et al. (2014). The fields represent the composition of melts derived from a range of potential sources (tonalites, metasediments, low- and high-K mafic rocks).

Although the granodiorite gneiss suite shows internal compositional heterogeneity, the two dated samples (KD-03, ZG-07) plotting unambiguously in the sanukitoid field and the remainder distributed across the overlap zone. The cumulative geochemical signature e.g., high Ba+Sr, LILE enrichment, elevated Mg#, low-to-moderate HREE fractionation, and dominantly $K_2O/Na_2O > 0.5$ supports a common sanukitoid parentage. The spread therefore reflects variable fractional crystallization and crustal assimilation rather than distinct source affiliations. Accordingly, the two dated samples are treated as geochemically unambiguous sanukitoids that constrain the timing of sanukitoid magmatism. Such rocks are widely interpreted as products of partial melting of mantle peridotite enriched by the addition of crustal-derived fluids or melts in a subduction zone setting. Subsequent evolution of these magmas was governed by fractional crystallization, with variable contributions from crustal assimilation during ascent. In the Laurent et al. (2014) source discrimination diagram, the samples plot predominantly in the low-K to high-K mafic rock field (Figure 8b), consistent with derivation from a mafic mantle source. Notably, the appearance of sanukitoid as mesosomes within migmatite outcrops likely reflects entrainment within a younger post-2545 Ma migmatitic episode, such that these rocks in the incorporated as mesosomes within a younger anatectic system. Ghanapur area preserve a dual identity: primary sanukitoid intrusive by origin, subsequently incorporated as mesosomes within a younger anatectic system.

The leucogranites represents the most evolved members of the suite, and although undated in this study, are interpreted as terminal products of crustal anatexis. Their high SiO_2 , stronger peraluminous character, low REE budget, low Mg# and pronounced negative Eu anomalies indicate derivation by partial melting of older tonalitic to granodioritic crustal material. The pronounced negative Eu anomalies, together with low Ba, Sr, and Ti, record a dominant control by residual and/or fractionating plagioclase (with K-feldspar and Fe-Ti oxides), whereas the marked enrichment in Rb, Th, and U reflects extreme late-stage fractionation. The elevated HREE profile in leucogranites is related to the presence of peritectic garnet, which is evident from petrological observation. Collectively, these features are consistent with high-temperature, fluid-absent (biotite-dehydration) anatexis of tonalitic to granodioritic crust during regional crustal thickening, thermal relaxation, and migmatization, which generated and mobilized these small-volume peraluminous melts (Harris et al., 1993; Clemens, 2003). In the Laurent et al. (2014) source-discrimination diagram, the samples plot in tonalite field (Figure 8b), consistent with derivation from

felsic to intermediate crustal protoliths. This is consistent with model in which small volume leucocratic melts segregate during thermal relaxation and migmatization of thickened crust, pointing to the waning stages of late Archean crustal thickening as most possible tectonic setting (Harris et al., 1993).

Geochronological constraints and tectonic implications

The zircon U–Pb ages define a prolonged late Archean magmatic and reworking history for the granitoid suite, spanning from 2697 ± 24 Ma to 2545 ± 25 Ma. The granitic gneisses yield crystallization ages of 2697 ± 24 Ma (TG-04) and 2607 ± 41 Ma (SR-01), whereas the granodiorite gneisses yield ages of 2560 ± 20 Ma (KD-03) and 2545 ± 25 Ma (ZG-07) (Figure 7). This age spread indicates that granitoid generation was not a single short-lived event but rather a protracted process involving repeated melt extraction, emplacement, and crustal reworking. Such stepwise granitoid development is typical of late Archean terranes, where crustal thickening, thermal reworking, and episodic magma addition operated together during craton assembly (Manikyamba and Kerrich, 2012; Jayananda et al., 2018).

Although widely used, the tectonic discrimination diagrams of Pearce et al. (1984) were developed primarily for post-Archean granitoids and should be applied to Archean granitoids with caution. The tectonic-affinity interpretations derived from these diagrams are therefore treated as suggestive rather than definitive, and the strongly peraluminous leucogranites in particular are interpreted with corresponding caution, as such compositions are not well suited to these discrimination schemes (Whalen and Hildebrand, 2019). Most importantly, differentiation can produce compositional trends that cross field boundaries, particularly from the volcanic arc granite (VAG) to within plate granite (WPG) field (Pearce et al., 1984). Nevertheless, the granitoids from Ghanapur area cluster mainly in the syn-collisional and volcanic-arc granitic fields, with few samples extending towards the within-plate granite field in the Pearce et al. (1984) discrimination diagram (Figure 9a-d). This distribution suggests a broadly convergent-margin affinity that evolved through time towards increasingly fractionated and internally reworked felsic magmatism. Such a pattern is consistent with Archean crustal growth involving subduction-related mantle metasomatism and subsequent melting, followed by magma hybridization, crystal fractionation, and variable contributions from remelting of older crustal material (Moyen, 2011; Laurent et al., 2014). The granitoid suite of the Ghanapur region therefore records a late Archean magmatic history spanning at least 150 Ma, reflecting a progression from early crustal melting

represented by the older granitic gneisses, through younger hybrid crust-mantle magmatism expressed by the granodiorite gneisses, to terminal high-temperature crustal anatexis preserved in the leucogranites.

The sanukitoid affinity of the granodiorite gneiss is however of particular importance because sanukitoids are widely interpreted as markers of crust-mantle interaction (e.g., Martin and Moyen, 2002; Laurent et al., 2014). It is worth noting that, sanukitoid magmatism in the EDC has been attributed to melting of an enriched mantle source following subduction along the EDC-WDC convergent margin, with slab breakoff commonly invoked as the trigger for asthenospheric upwelling and juvenile magma addition (Jayananda et al., 2020; Raju et al., 2024). The broadly synchronous occurrence of this magmatic episode across the craton at ca. 2570–2510 Ma (e.g., Jayananda et al., 2020; Dey et al., 2017) may, however, indicate melting of a previously enriched lithospheric mantle on a craton-wide scale rather

than a direct response to a single, spatially continuous convergent margin. The northeastern EDC presents a particularly important yet poorly constrained case in this context. The Ghanapur area is geographically distant from the proposed collision zone. Whether the sanukitoid magmatism documented here at ca. 2560–2545 Ma was generated through the same tectono-magmatic processes proposed for the other parts of the EDC remains unresolved. Alternatively, it may reflect a distinct history of mantle enrichment linked to the crustal architecture of the northeastern margin. Addressing this question will require further investigation of the Ghanapur granodioritic gneisses together with structural constraints on the deformation associated with the ca. 2560–2545 Ma magmatic event. Consequently, the northeastern EDC provides an important opportunity to evaluate the extent to which late Neoproterozoic mantle enrichment and sanukitoid-like magmatism varied across the Dharwar craton.

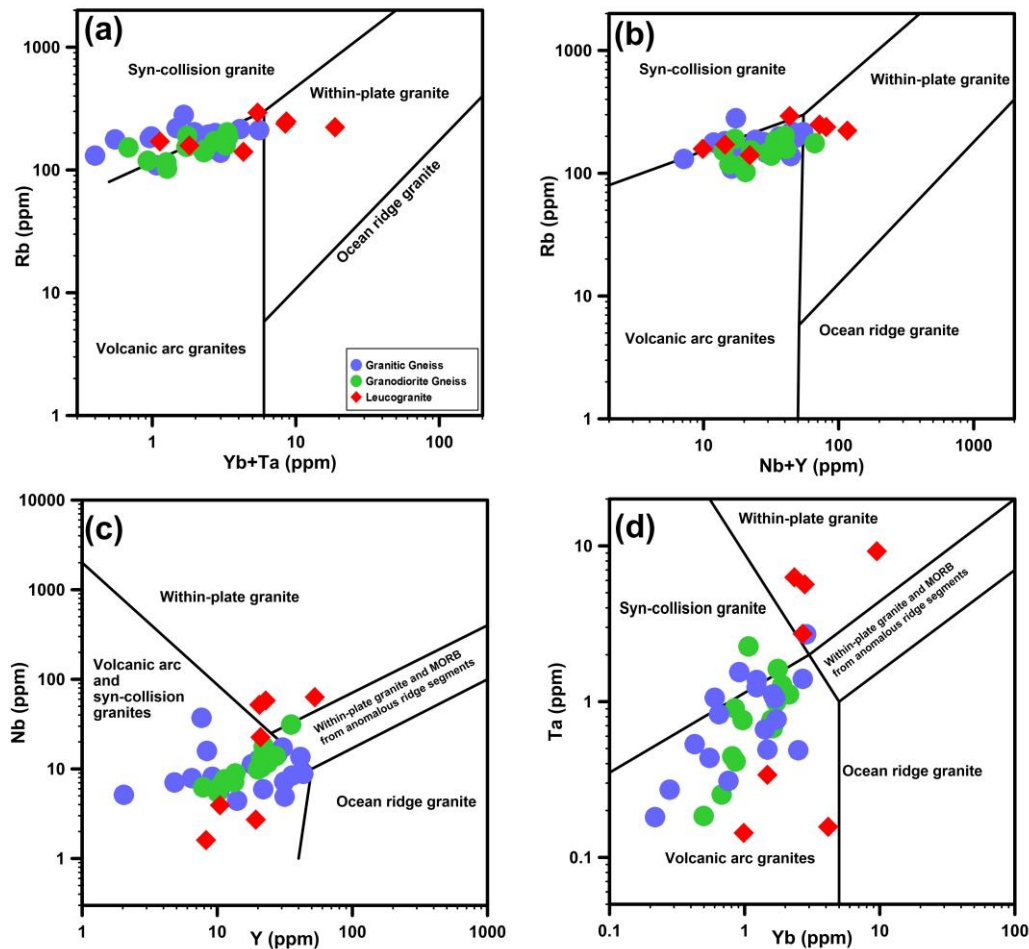


Figure 9. (a-d) Tectonic discrimination diagrams for granitoids based on trace element compositions after Pearce et al. (1984).

CONCLUSIONS

- (i) The Ghanapur area preserves a coherent record of the late Archean crustal growth, reworking, and progressive crustal maturation within the EDC.
- (ii) The studied rocks are classified into three suites on the basis of their field relationships, petrographic features, and geochemical characteristics: granitic gneisses, granodiorite gneisses, leucogranites.
- (iii) The older granitic gneisses (2697–2607 Ma) record early crust-forming events dominated by melting of heterogeneous crustal protoliths, including metasedimentary, tonalitic and high-K mafic sources.
- (iv) The younger granodiorite gneiss (2560–2545 Ma) show sanukitoid affinity, indicating derivation from melting of a mantle wedge metasomatized by subduction-derived fluids and melts.
- (v) The leucogranites represent terminal high-temperature crustal anatexis, formed as small-volume late anatectic melts.
- (vi) The geochemical and geochronological constraints are consistent with a broadly convergent-margin setting in which repeated episodes of crustal melting, mantle wedge metasomatism, and thermal reworking contributed to late Archean crustal evolution in the Ghanapur region over a protracted period of ~150 Ma.

Acknowledgements

The authors thank Dr. Prakash Kumar, Director, CSIR-National Geophysical Research Institute (CSIR-NGRI), for granting permission to undertake this work. We thank the chief editor Dr. O.P. Pandey for efficient handling of the manuscript and the editorial comments. N.K. acknowledges financial support from the Department of Science and Technology (DST) under the Women Scientist Scheme-A (SR/WOS-A/EA-41/2017). N.K. is thankful to Dr. J.K. Tomson for his support during the XRF analysis at NCESS, Thiruvananthapuram, India. The authors are grateful to Dr. Bulusu Sreenivas for his guidance and support during the field investigations. We also thank the three anonymous reviewers for their constructive comments, which significantly improved the quality of the manuscript. This study forms a part of the Ph.D. research work of the first author.

Author Credit statement

Namrata Khadke: Conceptualization, methodology, investigation, formal analysis, data curation, visualization, funding acquisition, software, writing original draft, writing

review & editing. Sarbajit Dash: Validation, writing, review & editing. T. Vijaya Kumar: Formal analysis. E.V.S.S.K. Babu: Supervision, writing, review & editing. Y.J. Bhaskar Rao: supervision and resources

Data availability

The datasets generated and/or analyzed during the current study are available from the corresponding author upon reasonable request.

Compliance with Ethical Standards

The authors declare no conflict of interest. The manuscript complies with all applicable copyright requirements and ethical standards for publication.

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Pressure dynamics controlling crude oil migration for reservoir management and exploration in Niger Delta, Nigeria

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ABSTRACT

Pressure conditions are required in reservoir operations management. This is essential in the adjustment of migration and production strategies. The study investigates the relationship between fluid flux and static pressure across multiple wells in a Niger Delta reservoir using well log and pressure data. Fluid flux and pressure gradients were obtained from modified Darcy's Law, and these were used to generate plots for oil, gas, and oil/water phases. A threshold pressure of approximately 4,899 psi was identified beyond which fluid flux increased sharply, reaching a maximum of 1,400 ft³/day. Below this threshold pressure, fluid mobility was minimal, which indicates an insufficient energy to overcome capillary barriers. The optimal pressure suggests enhanced bulk fluid transmissibility, where both metrics exhibited a sharp decline, stabilizing at lower positive values until approximately 4800 psi. A critical pressure range of 4800 psi to 5050 psi was identified as a pressure range for optimal fluid flux. The major fluid encountered across the entire pressure spectrum was identified as oil, with isolated occurrences of gas and oil/water mixtures. In the present work, we observed that there is existence of a pressure-controlled migration regime and emphasize the importance of maintaining reservoir pressure above threshold values to enhance the oil production. The results obtained would provide insights for designing enhanced oil recovery (EOR) strategies and maintaining long-term reservoir performance above the present traditional concepts of oil production.

Keywords: Threshold Pressure Gradient (TPG), Crude oil migration, Darcy's Law, Seepage velocity, Reservoir management, Niger Delta basin

Nomenclature

Symbol	Description	Units
q	fluid flowing velocity	cm/s
$\emptyset$	Porosity	Fraction or %
k	Permeability	mD
μ	viscosity of the flowing fluid	centipoise
dp/dx	pressure gradient in the direction of flow	atm/cm
x	distance in the direction of flow	Cm
ΔP	total pressure drops	Psi
P_a	initial pressure	Psi
P_b	final pressure	Psi
l_a	initial depth	Ft
l_b	final depth	Ft
q	fluid flux	ft ³ /day
v_s	seepage velocity	m/s

INTRODUCTION

Petrophysics integrates geological, physical, and chemical properties of reservoir rocks to evaluate parameters such as porosity, permeability, volume of shale, and fluid saturation for assessing reservoir productivity (Shah et al., 2021; Shalaby et al., 2023). The well log interpretation, core analysis, and pressure monitoring help petrophysical evaluation to provide critical inputs for reservoir management. When accurate and timely data are provided, petrophysics enables operators to make useful decisions that optimize production rates, reduce operational risks, and improve total reservoir management (Okoro et al., 2022; Esiri et al., 2024; Iriogbe et al., 2024). The significance of

enhanced reservoir management lies in its ability to firstly, improve reservoir understanding and characterization, thereby enabling more accurate estimation of reservoir properties and production potential. This is crucial in mature fields where remaining reservoirs may be complex to quantify using conventional methods. Secondly, advanced monitoring techniques enable real time monitoring of reservoir performance, allowing operators to promptly detect and respond to changes in reservoir situations. This can help to identify issues such as gas coning, water breakthrough, or production anomalies, thus modifying potential production losses and extending the reservoirs life (Satter and Iqbal, 2016; Wu et al., 2023).

Traditional reservoir characterization methods, including core analysis and well logging, have long been used to evaluate reservoir properties. However, core samples are often limited in spatial coverage and may not adequately represent reservoir heterogeneity (Aminzadeh and Dasgupta, 2013; Ganguli and Dimri, 2023; Onita and Ochulor, 2024). Similarly, although well logs provide continuous subsurface information, their interpretation can be affected by clay content and other minerals, leading to uncertainties in porosity and fluid saturation estimates (Bhowmick et al., 2024). These limitations have encouraged the development of more advanced technologies for improved reservoir characterization and management (Ganguli and Dimri, 2023; Onita and Ochulor, 2024).

Enhanced reservoir monitoring supports the implementation of advanced recovery techniques, such as enhanced oil recovery (EOR) methods. By providing comprehensive insight into reservoir behaviour, these techniques can be tailored to the specific conditions of the reservoir, maximizing recovery efficiency and minimizing negative environmental impact (Ahmed, 2024). This could be done by integrating some information from the petrophysical parameter, such as pressure into Machine learning and AI, which are vital tools that can be used to automate data interpretation and optimize reservoir management strategies (Krasnov et al., 2017; Yahya et al., 2025; Singh et al., 2025).

The existing studies have examined generalized reservoir pressure behaviour without quantitatively resolving pressure thresholds governing hydrocarbon mobility under heterogeneous Niger Delta conditions (Vaferi and Eslamloueyan, 2015; Salami and Nton, 2024). The existing literature has not adequately quantified threshold pressure, defined optimum pressure ranges, or established maximum pressure limits required for effective migration of crude oil across capillary barriers and fault zones in Niger Delta. Most prior investigations either focus on generalized pressure regimes without differentiating threshold from maximum migration pressures or constrain their research to specific lithologies and basin conditions (Gao and Flemings, 2016; Satter and Iqbal, 2016; Hong et al., 2018). Present work exposes a critical knowledge gap, particularly in the Niger Delta, where heterogeneity in pore throat sizes, capillary forces, and reservoir connectivity influence the hydrocarbon mobility. It provides a quantitative evaluation of threshold and optimum pressures governing hydrocarbon migration, offering a practical framework for improving reservoir pressure management in heterogeneous basins. It aims to emphasize the significance of pressure in operational petrophysics and its role in enhancing reservoir management,

thereby contributing to more efficient and sustainable oil and gas production practices. The originality of the research is based on three main contributions. First, this paper provides field-specific quantification of pressure thresholds influencing crude oil migration in the Niger Delta's Freeman Field, which has not been done before. Precisely, a threshold pressure of about 4,899 psi, an ideal pressure range of 4,800–5,050 psi, and a maximum pressure of roughly 4,899 psi, are estimated and deduced in light of the local reservoir planning.

In order to identify crucial migration windows throughout the reservoir system, a multi-well, multi-phase pressure-flux design that integrates data from the eight wells spanning oil, gas, and oil/water phases was created. A pressure interval linked to capillary breakthrough or fracture pathway activation has been identified as the crucial transition zone around 4,800–5,050 psi. This finding has significant operational consequences for pressure maintenance and the design of enhanced oil recovery (EOR) strategies. Though often linked to standard recovery methods, EOR plays a broader role here. As pressures drop over time in aging Niger Delta reservoirs, output becomes unstable. Gas begins to cone, water arrives early, and native drive forces weaken - each undermining how much oil can be extracted. These issues also cut short the operational lifespan of wells (Ahmed, 2024). Starting at roughly 4,899 psi, fluids begin moving reliably through the formation, below this level, flow stalls. Maintaining pressures between 4,800 and 5,050 psi, supports efficient movement across the reservoir zone.

FIELD GEOLOGICAL SETTINGS AND CHARACTERISTICS

The field under review is placed inside the offshore portion of the Niger Delta. The Niger Delta is located inside the Gulf of Guinea at the border of West Africa between 3° and 6°N and 5° and 8° E, close to the west coast of Nigeria. The Niger Delta sits on the southern edge of the Benue trough, corresponding to a trapped branch of a triple rift. It is one of the most important and prolific regressive deltas for oil production (Uchegbulam and Ayolabi, 2014; Abiola and Obasuyi, 2019). It has a sub aerial region of approximately 75,000 km², a complete region of 300,000 km² and a sediment fill of 500,000 km² (Reijers, 2011). The Niger Delta basin known as the Niger Delta province, is a vastly productive one and it is of excessive marketable worth to Nigeria. The basin includes merely one petroleum system, identified as the Tertiary Niger Delta (Akata Agbada) petroleum system. The Niger Delta's sedimentary wedge is distributed across five depobelts with the earliest placed offshore and the oldest located furthest away inland (Reijers, 2011; Jonathan and Charles, 2025). These depobelts include,

offshore depobelts, Central Swamp, Northern Delta, Coastal Swamp and Greater Ughelli. Each depobelt is bounded inward by growth faults and seaward by huge counter-confined faults that come after sea-ward depobelt (Reijers, 2011; Jonathan and Charles, 2025). The reservoir units analyzed in this study are predominantly associated with the Agbada Formation, characterized by interbedded sand – shale sequences typical of the Niger Delta. The sand bodies serve as the primary hydrocarbon reservoirs, while the interbedded shales, act as flow barriers and seals. This lithological heterogeneity significantly influences pressure distribution and fluid mobility across the wells.

Stratigraphy of the Niger Delta Basin

The tertiary sequence established in the Niger Delta consists, in increasing order, of the Akata, Agbada, and Benin formations (Figure 1). The strata formed an appraised depth of 8,535 m (28,000ft) near the centre of the deposit in the dominant part of the delta. The Akata is characterized by a

uniform growth of shale, generally dark grey, while in some places, it is silty or sandy, particularly in the upper part of the formation (Schlumberger, 1985), accordingly, the top of the Akata Formation provides an economic base for oil and gas exploration.

However, gas may be present at deeper levels due to high pressures. The Agbada Formation is a succession of interbedded sandstones and shales, containing sandstone reservoirs responsible for gas and oil production in the Niger Delta (Ayuk et al., 2022; Adagunodo et al., 2022). The sandstones are medium to clean, fine-grained, and locally rich in limestone, glauconite, and shells. The shale is medium to dark grey, consolidated, and locally contains glauconitic mud (Doust and Omatsola, 1990). The Benin Formation consists of massive, freshwater bearing, highly porous sandstones with locally thin layers of fluvial shale. The sandstone consists mainly of quartz and potash feldspar with minor amounts of plagioclase (Harry et al., 2022).

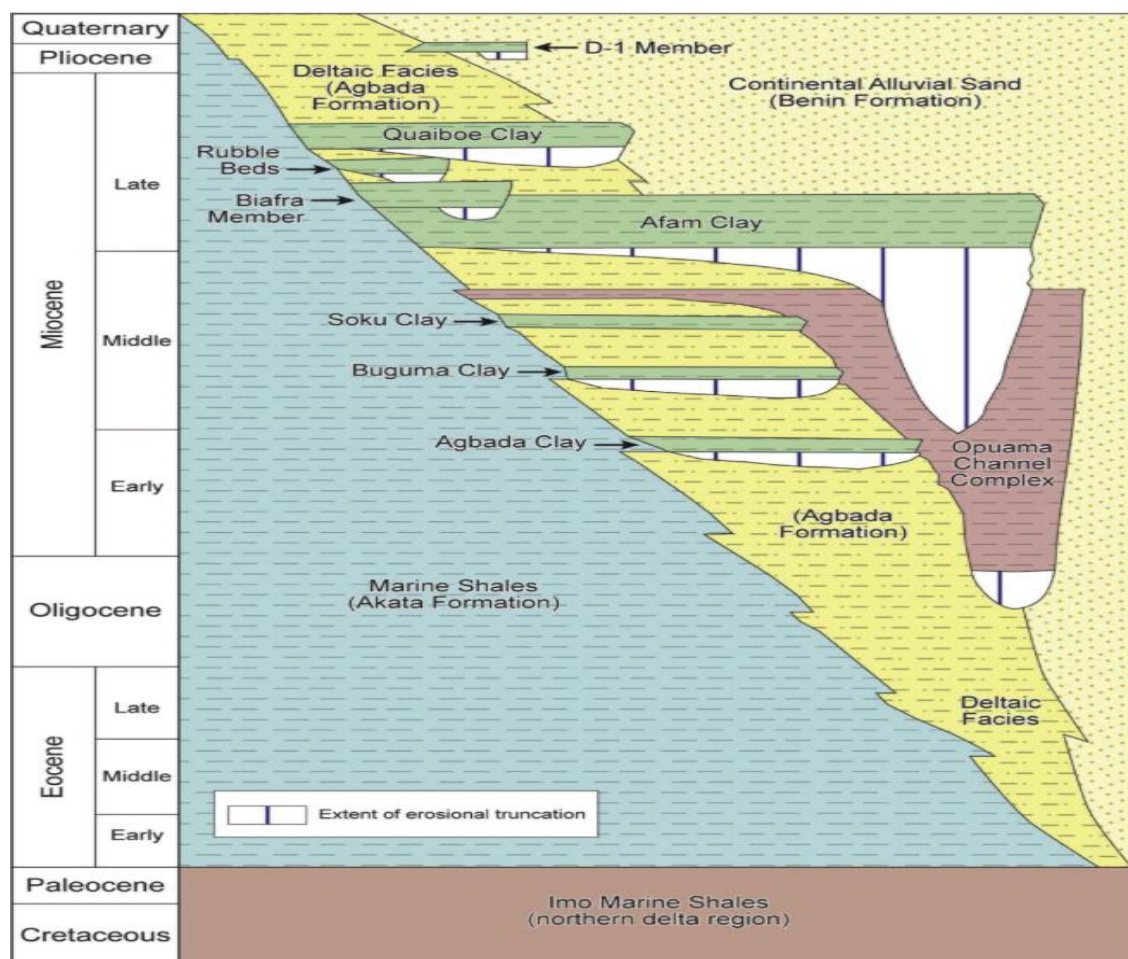


Figure 1. The three formations of the Niger Delta Stratigraphic column (Reijers, 2011)

METHODOLOGY

Data and study area

Eight wells were used in this work, the wells which aligned in a northwest direction towards the study area. Shell Petroleum Development Company of Nigeria Limited, through the Nigeria Upstream Petroleum Regulatory Commission, Lagos, Nigeria, donated the datasets used in this study. The log of the wireline for the wells includes, density, gamma ray, neutron logs, resistivity, sonic logs in LAS format and static pressure data (Table 1). The analysis was implemented in a modular Python script (compatible with IPython/Jupyter). Users can apply the code by inputting their own datasets in the modelled format.

Table 1 shows the wire-log data of all the wells used for this study. The GR log was used for the Well log-based characterization for sand-shale identification, neutron-density cross plot for porosity, and resistivity logs for fluid delineation. The petrophysical parameter was used for the estimating the average porosity of 20%, permeability estimation and fluid saturation. The lithological description of Agbada Formation reservoir sands, correlated across the eight Freeman Field wells but 3D seismic data were not available for this study. This restraint is clearly stated in a new uncertainty subsection, where the RPT analysis cited as a benchmark for future seismic integration work (Datta Gupta et al., 2012)

Theoretical framework

Reservoir Description

The available wireline log suite, which includes gamma ray (GR), resistivity, sonic (DT), neutron porosity (NPHI), and bulk density (RHOB) logs, was used to characterize the Freeman Field reservoir. The productive reservoir zones within the Agbada Formation were identified by using GR logs to separate sand from shale intervals based on deflections toward low API values. Effective porosity was estimated and fluid types were identified using neutron-density cross-plot analysis; gas-bearing intervals were indicated by the crossover between the neutron and density curves. The Archie equation and resistivity logs were utilized to distinguish hydrocarbon-bearing zones from water-saturated intervals. The reservoir sands have an average effective porosity of about 20%, which is in line with values published for the Niger Delta's Agbada Formation sandstones (Kong and Chen, 2018).

Permeability estimations were made using existing porosity-permeability transforms for Niger Delta sandstones. The reservoir sands are medium to fine-grained, reasonably clean (low shale content), and laterally continuous across the eight

Freeman Field wells, as evidenced by GR log correlation. These properties support the use of Darcy's Law as the principal flow model in this investigation (Datta Gupta et al., 2012). It is noted that no 3D seismic data were available for this investigation; this is a restriction because seismic data would further constrain structural geometry and fault connectivity. Future work should include rock physics template (RPT) analysis of seismic and well log data, as part of an effort to provide a more complete reservoir characterisation (Datta Gupta et al., 2012; Chakraborty et al., 2024).

The threshold, optimum, and maximum pressures for crude oil migration were determined by applying Darcy's Law to model fluid flow through porous media. The flow rate was expressed as a function of permeability, fluid viscosity, and pressure gradient. The threshold pressure is the minimum pressure required to overcome capillary forces and initiate flow; the optimum pressure range ensures efficient migration without inducing rock fracturing; maximum pressure is the limit beyond which nonlinear flow and mechanical failure of the reservoir occur. This approach integrates petrophysical properties, static pressure data, and capillary pressure information following some other techniques (Upadhyay et al., 2023; Mkpese et al., 2024; Nabawy, 2025); Fjaer et al., 2021

Darcy's Law describes the flow of fluids through porous media and explains the migration of hydrocarbons in petroleum reservoirs (Kuuskraa, 1982). This law, in its simple differential form is shown in Equation 1:

$$q = -\frac{k}{\mu} \frac{dp}{dx} \quad (1)$$

where q = apparent fluid flowing velocity, cm/s; k = permeability of the porous rock, Darcy;

μ = viscosity of the flowing fluid, centipoise; dp/dx = pressure gradient in the direction of flow, atm/cm; x = distance in the direction of flow, always positive, cm

In this study, the term "modified Darcy's law" does not imply the incorporation of non-Darcy or turbulent flow corrections. Rather, it refers to the practical adaptation of Darcy's equation to field scale reservoir conditions using discrete pressure measurements across well intervals and multiple fluid phases (oil, gas and oil/water systems). The formulation integrates pressure differences, depth variations and fluid properties obtained from well log and pressure datasets to estimate fluid flux under static reservoir conditions. The

negative sign in Darcy's law signifies that the direction of fluid flow is opposite to the direction of increasing pressure (Lyons, 2010; Fanchi, 2010). From an exploration point of view, integrating Darcy's law into basin modeling and reservoir simulations, allows geoscientists to better predict hydrocarbon accumulations and assess reservoir quality (Zhang et al., 2024). It helps identify the most probable migration routes and hydrocarbon traps while also determining whether a seal is effective in preventing further movement of hydrocarbons (Satter and Iqbal, 2016; Sunmonu et al., 2016; Akroun et al., 2021; Zhang et al., 2024).

Equation 2 depicts the total pressure drops under a steady incompressible fluid flow, as

$$\Delta P = \frac{\mu q \Delta L}{k} \quad (2)$$

$$q = \frac{k}{\mu} \left(\frac{P_b}{l_b} - \frac{P_a}{l_a} \right) \quad (3)$$

where q = Fluid flux, k = Permeability of the fluid, μ = Viscosity of the fluid, P_a = Initial pressure, P_b = Final pressure, l_a = Initial Depth, l_b = Final Depth

These expressions formed the basis for calculating pressure gradients and fluid flux for each well and fluid phase (Zhang et al., 2019; Wang et al., 2020). In the context of hydrocarbon exploration and production, it is crucial to understand how oil, gas, and water move through reservoir rocks (Yang et al., 2022). Equation 3 can be broken down to understand its components and their significance in hydrocarbon exploration. The volumetric flow rate (seepage velocity/ fluid

flux), q represents the seepage velocity or the volumetric flow rate of the fluid (oil, gas or water) through a unit cross-sectional area of the porous medium (Wang et al., 2020; Yang et al., 2022; Ahmed et al., 2023; Zhang et al., 2024).

The permeability, k denotes the absolute permeability of the porous medium (Yang et al., 2022). Permeability is directly influenced by the rock's pore structures, including pore size, type, and connectivity, as well as the presence of fractures (Qiao et al., 2021; Yang et al., 2022; Yu et al., 2022; Gao et al., 2024). Similarly, Fluid viscosity, μ represents the dynamic viscosity of the fluid (Yang et al., 2022). Viscosity is a measure of a fluid's resistance to flow (Yang et al., 2022; Gao et al., 2024). The significance of fluid viscosity in hydrocarbon exploration is that, the lower viscosity fluids (light oil, gas) tend to migrate and flow more easily through porous media than highly viscous fluids (heavy oil) (Yang et al., 2022; Gao et al., 2024; Zhang et al., 2024). Fluid properties such as viscosity were assumed based on typical values for Niger Delta reservoir fluids due to the unavailability of detailed PVT data. These assumptions are considered reasonable for comparative analysis of pressure–flux relationships.

$\frac{P_b - P_a}{l_b - l_a}$ represents the pressure gradient across the porous medium, where $P_b - P_a$ is the pressure difference between two points (b and a), and $l_b - l_a$ is the distance between those points.

Table 1. Log data provided for this study

WELLS	WIRE LOGS					
	GR	RES	DT	NPHI	RHOB	PRESSURE
FREEMAN 1	√	√	√	√	√	√
FREEMAN 2	√	√	√	√	√	√
FREEMAN 2ST	√	√	√	√	√	√
FREEMAN 3	√	√	√	√	√	√
FREEMAN 3ST	√	√	√	√	√	√
FREEMAN 4ST	√	√	√	√	√	√
FREEMAN 8	√	√	√	√	√	√
FREEMAN 5	√	√	√	√	√	√

Table 2. Fluid property values used for this study (PVT properties)

Property	Values
Oil viscosity	1.5 cp (at reservoir conditions)
Oil density	0.85 g/cc
API gravity	32°
Gas viscosity	0.018 cp
Water viscosity	0.5 cp
Bo	1.2 RB/STB

RB=Reservoir Barrel, STB=Stock Tank Barrel, Bo=Oil Formation Volume Factor, PVT= Pressure, Volume, Temperature

It signifies the driving force behind fluid flow (Yang et al., 2022; Huang et al., 2023). Fluids tend to flow from areas of high potential to low potential (Huang et al., 2023). This study focuses on pressure – flux relationships and does not explicitly incorporate spatial modeling or well interference effects. However, the wells were analyzed collectively to capture general reservoir behavior. In Table 2, the PVT properties are given, which shows the fluid property values used for this work. The values were estimated from published Niger Delta fluid property correlations (Standing, 1947; Ahmed, 2024) in the absence of laboratory PVT data, and this is explicitly noted in the table caption.

Analytical Workflow

This study adopted a computational and data driven approach to investigate the pressure behaviour and fluid dynamics within the hydrocarbon reservoirs of the Niger Delta. Pressure and flow rate data acquired from multiple wells, were pre-processed to remove incomplete records and standardized for consistency. Key parameters, such as reservoir pressure (P), pressure difference (ΔP), pressure gradient ($\Delta P/\Delta L$), fluid type, and volumetric flow rate (q), were compiled into structured datasets for analysis. The pressure gradient was determined by dividing the pressure difference across reservoir sections by the measured distances between sampling depths. Seepage velocity (v_s) was computed using the relationship $v_s = q/\phi$, where q represents the volumetric flow rate and ϕ is the porosity, which was assumed to be 20% in alignment with reported average porosity values for Niger Delta sandstones (Adagunodo et al., 2017; Kong and Chen, 2018).

Exploratory data analysis was performed to evaluate the relationships between fluid flux and pressure variables. Scatter plots and line plots of q against ΔP , P, and $\Delta P/\Delta L$ were generated both on linear and logarithmic scales for individual wells and for the aggregated dataset comprising all wells. To identify threshold and optimum pressure intervals, pressure data were grouped in 100-psi bins; mean fluid-flux values and their standard deviations were calculated within each bin to reveal pressure zones exhibiting maximum flow rates.

Statistical analysis was employed to quantify reservoir behaviour across fluid types. Summary of the statistics including standard deviation, mean, minimum, and maximum values of seepage velocity and pressure were calculated for water, oil, and gas. Linear regression analysis was performed to describe the relationship between pressure drop and fluid flux. All computations and graphics were produced with Python based libraries including Matplotlib

and Seaborn. The resulting scatter and faceted plots facilitated comparison of flow characteristics among wells and fluid types. The final dataset, enriched with derived parameters and fluid classification codes, was archived for interpretation and validation. All results were exported as high-resolution figures for reporting. It is important to note that this study utilizes static reservoir pressure data, which represents equilibrium conditions within the reservoir. While dynamic pressure (drawdown) is critical during production, static pressure is appropriate for evaluating hydrocarbon migration and pressure-controlled flow behavior. The effect of dynamic pressure conditions is considered a limitation and recommended for future investigation.

Uncertainty Analysis

This study contains a number of inherent sources of uncertainty, which are acknowledged as follows. The first source of input data uncertainty is the variance in static pressure readings among the eight wells, which may be caused by variations in gauge precision, measurement depth, and reservoir compartmentalization, rather than just actual pressure variation. Second, based on published average values for Niger Delta Agbada Formation sandstones, a consistent porosity of 20% was assumed for all wells and reservoir intervals (Kong and Chen, 2018). Seepage velocity was recalculated using porosity values of 15% and 25%, which correspond to the lowest and upper boundaries of reported Niger Delta reservoir porosity, in order to evaluate the sensitivity of this assumption. The resulting seepage velocities differed from the base case by about $\pm 25\%$, which has an impact on the absolute magnitude of computed flux values, but does not change the identification of the threshold pressure range. Third, laminar, single-phase flow over a homogeneous porous media is assumed when applying Darcy's Law. Deviations from this assumption increase model uncertainty that cannot be adequately quantified without core measurements or well test data, especially in fractured or heterogeneous zones and in multi-phase flow situations. Near transition areas like the oil-water contact (OWC) and gas-oil contact (GOC), mixed-phase signals can appear misleading. This happens because logs assign phase types through coded outputs. Pressure-threshold findings rely on data showing large variations across pressure bins, particularly between 4800 and 5050 psi. Such variability reflects both measurement uncertainty and changes in reservoir flow behaviour. Although these fluctuations do not alter the identification of the threshold pressure range, operational application of the threshold values should consider the associated uncertainty.

RESULTS AND DISCUSSION

Determination of maximum pressure ($P_{\max}$), threshold pressure (P_{thrsh}), and pressure range (P_{range}) is required to enhance the migration and volume of hydrocarbon (oil and gas) in complex reservoirs. The exact pressure (not just pressure drop) was identified yielding maximum fluid flux. The data were grouped by pressure, and a threshold pressure (P_{thrsh}) of approximately 4899 psi was identified at a fluid flux value of 1400 ft³/day, as illustrated in Figure 2.

Determination of maximum pressure ($P_{\max}$)

The relationship between reservoir pressure and fluid flux across all wells within the study area was illustrated in Figure 2. The plot depicts fluid flux as a function of reservoir pressure (psi), with each well represented by distinct color-coded markers and lines for clarity. A notable feature in the graph is the red dashed vertical line, which identifies the maximum fluid flux of 1,400 ft³/day occurring at a threshold pressure (P_{thrsh}) of approximately 4899 psi. This critical point is further highlighted with a red marker to emphasize its significance.

The results demonstrate that at pressures below approximately 4,000 psi, fluid flux remains relatively low across most wells, often fluctuating near zero and occasionally showing negative values. These low or negative values may be attributed to insufficient reservoir pressure to

overcome capillary entry pressures, resulting in limited hydrocarbon mobility and, in some cases, potential backflow within certain well sections. As reservoir pressure increases toward the range of 4500 to 4900 psi, fluid flux begins to rise more noticeably, indicating enhanced hydrocarbon migration. The identified threshold pressure (~4899 psi) is interpreted as the capillary entry pressure, required to overcome pore throat resistance within the reservoir. At this pressure, hydrocarbon fluids gain sufficient energy to displace resident fluids and migrate through constricted pore networks. Additionally, this threshold may correspond to localized capillary breakthrough or activation of high-permeability pathways such as micro-fractures.

This trend suggests the presence of a threshold pressure; the minimum pressure required to initiate and sustain effective hydrocarbon flow through the reservoir's pore network. The occurrence of maximum flux at 4899 psi signifies that this threshold was reached or slightly exceeded at this point. Wells such as Freeman-2ST3 and Freeman-4ST1, exhibit pronounced spikes in flux within this critical pressure range, potentially reflecting localized zones of higher permeability or favourable structural features that promote fluid migration (Akrouf et al., 2021; Zhang et al., 2024). They also highlight the importance of maintaining reservoir pressures above critical thresholds to sustain production efficiency in complex reservoirs (Dolson, 2016).

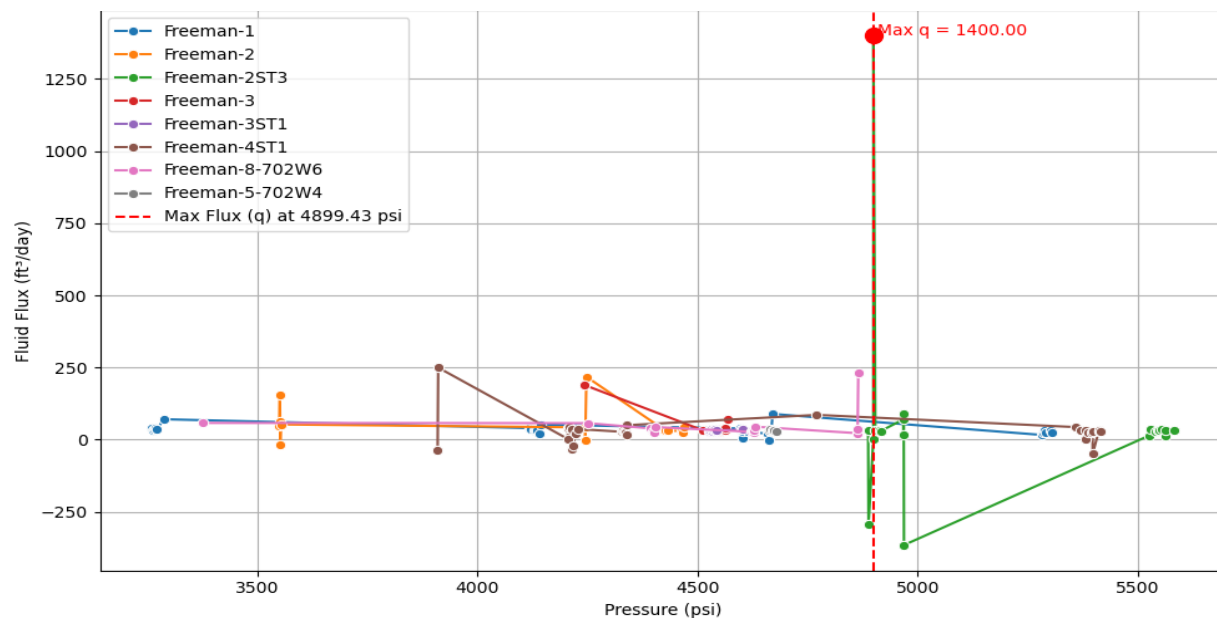


Figure 2. The plot showing relationship between reservoir pressure and fluid flux for all wells showing maximum flux at 4899 psi.

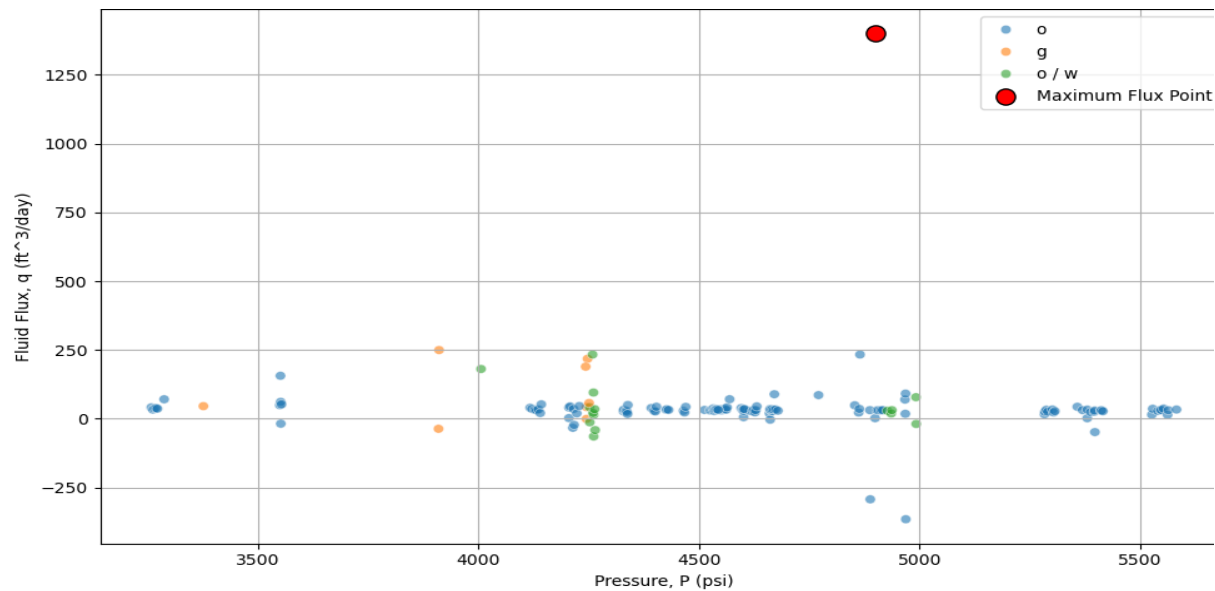


Figure 3. The plot showing maximum fluid flux versus pressure. Maximum fluid flux 1,400 ft³/day can be noted at 4899 psi pressure.

Relationship between reservoir pressure and fluid flux

The relationship between reservoir pressure and fluid flux for various fluid phases across the study area is shown in Figure 3. The plot shows fluid flux (ft³/day) on the vertical axis versus pressure (psi) on the horizontal axis, with data points color-coded by fluid type oil (“o”), gas (“g”), and oil-water mixtures (“o/w”). A prominent red marker highlights the maximum fluid flux point, recorded at approximately 1,400 ft³/day and occurring at a pressure of 4899 psi. This critical point is further emphasized by a red notation labeled “Maximum Flux Point”. The distribution of data points reveals that at pressures below 4000 psi, fluid flux across all phases remains relatively low, typically under 250 ft³/day, and in some instances, negative values are observed. These negative fluxes reflect localized backflow or pressure insufficiencies unable to overcome the capillary entry pressures of the reservoir matrix (Yang et al., 2022).

As reservoir pressure increases within the range of 4500 to 4900 psi, a noticeable increase in fluid flux is observed, with the oil phase (blue points) showing the most prominent response. This suggests that the onset of significant hydrocarbon migration occurs within this pressure interval, reaching a peak at 4899 psi. Beyond this critical pressure, no further substantial increases in fluid flux are recorded, indicating a plateauing behavior likely due to reservoir heterogeneities such as permeability barriers or capillary end effects (Adagunodo et al., 2017). This finding has significant

operational implications; maintaining reservoir pressures at or above the identified threshold (4899 psi) is essential for sustaining hydrocarbon flow and maximizing recovery (Santra et al., 2021).

Determination of threshold pressure, (P_{thrsh})

The relationship between pressure and mean fluid flux across all wells was investigated, revealing significant variability in fluid flow behaviour at specific pressure ranges. From approximately 4800 psi to 5600 psi, mean fluid flux demonstrated a general increasing trend, peaking at 4899 psi threshold pressure (P_{thrsh}). A sharp decline followed this peak, with flux values stabilizing positively to about 4800 psi. A notable increase in mean fluid flux was observed around 4900 psi, which was then succeeded by a dramatic decrease in negative values. This pressure range, between 4800 psi and 5050 psi, was characterized by a substantial standard deviation, indicative of high variability and uncertainty in fluid flux measurement (Figure 4). Beyond 5100 psi, the mean fluid flux gradually recovered to positive, relatively stable values. This observed high variability at approximately 4900 psi, suggests a critical pressure range where the hydrocarbon fluid flow dynamics exhibit significant and potentially unpredictable changes, transitioning from positive to negative flow, a phenomenon that warrants further investigation to understand underlying geological or operational factors.

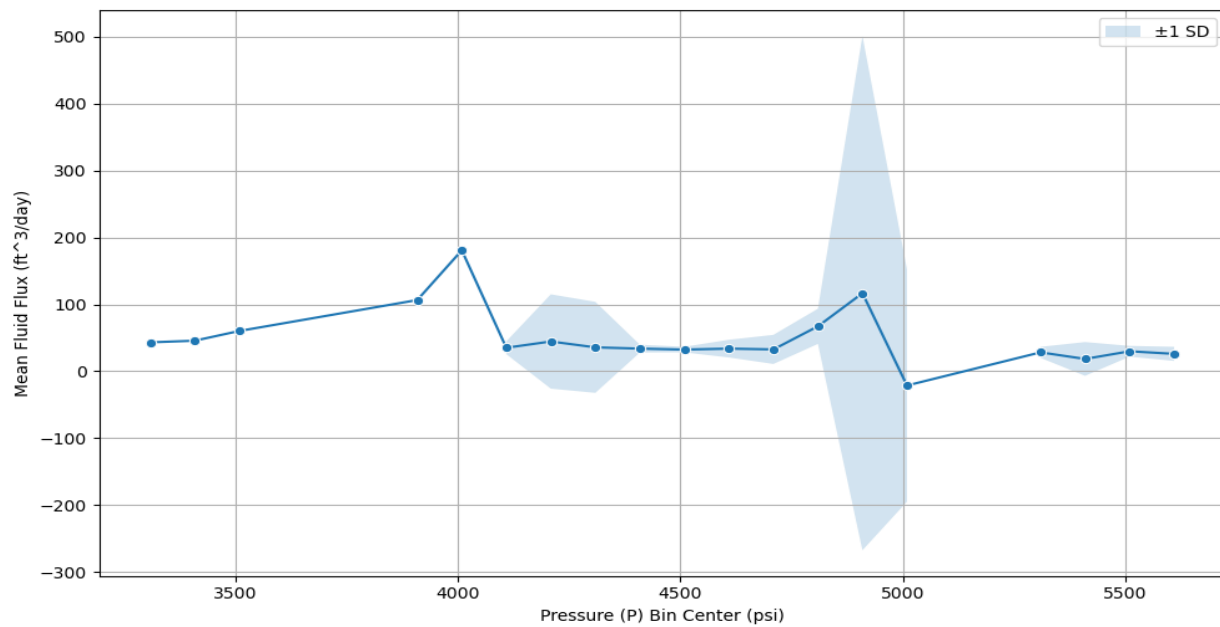


Figure 4. The plot between mean heat flux versus threshold pressure, showing the threshold pressure (P_{thrsh}) is 4899 psi.

Comparative analysis

The 4,899 psi P_{max} value corresponds to a pressure gradient of approximately 0.44–0.47 psi/ft at the sampled depths (approximately 10,400–11,000 ft), which is consistent with the normal hydrostatic gradient range (0.43–0.47 psi/ft) reported for the Agbada Formation in the central Niger Delta (Adagunodo et al., 2017; Akroun et al., 2021). This suggests that maximum hydrocarbon mobility in the Freeman Field occurs at near-normal hydrostatic conditions, rather than at over-pressured states, which has important implications for pressure maintenance design. Fields in the shallow offshore Niger Delta with similar settings have reported production threshold pressures in the 3,500–5,200 psi range (Harry et al., 2022), broadly consistent with our P_{thrsh} of 4899 psi.

Analysis of fluid flow system

The analysis of fluid flow across the study area revealed distinct pressure dependent regimes and a critical threshold range influencing overall system behavior. Mean fluid flux, derived from bin averaged pressure data across all wells (Figure 4), demonstrated an initial increase from approximately 3300 psi to a peak of around 4000 psi, indicating optimal bulk fluid transmissibility within this pressure window. Beyond this peak, flux values exhibited a sharp decline, stabilizing in positive territory until approximately 4800 psi. Critically, the pressure range between 4800 psi and 5050 psi, represented a transition zone

characterized by a dramatic shift in mean fluid flux, from positive to negative values, coupled with an exceptionally wide standard deviation. This substantial variability signifies a highly heterogeneous and unstable flow regime within this specific pressure interval. Further detailed investigation into individual well performance (Figure 5a), corroborated this finding, pinpointing a maximum seepage velocity of 7000 ft/day for well Freeman-2ST3 at a pressure of 4899 psi. This singular, high magnitude fluid event, occurring precisely within the identified threshold pressure range, strongly suggests the activation of a highly conductive pathway, potentially indicative of capillary breakthrough of a sealing formation or the propagation of a significant fracture network enabling rapid fluid migration. Equally, other wells within the same pressure range exhibited widely disparate seepage velocities, including instances of near-zero or negative flow, further emphasizing the localized and highly variable nature of fluid transport at this critical pressure.

Beyond 5050 psi, the mean fluid flux gradually stabilized into a new positive, albeit lower, equilibrium regime, with reduced variability. These findings collectively establish the pressure range 4800 to 5050 psi as a critical threshold, where reservoir scale fluid flow transitions fundamentally, likely governed by localized geological features and exceeding specific geomechanical or petrophysical thresholds for fluid mobility (Bai et al., 2020).

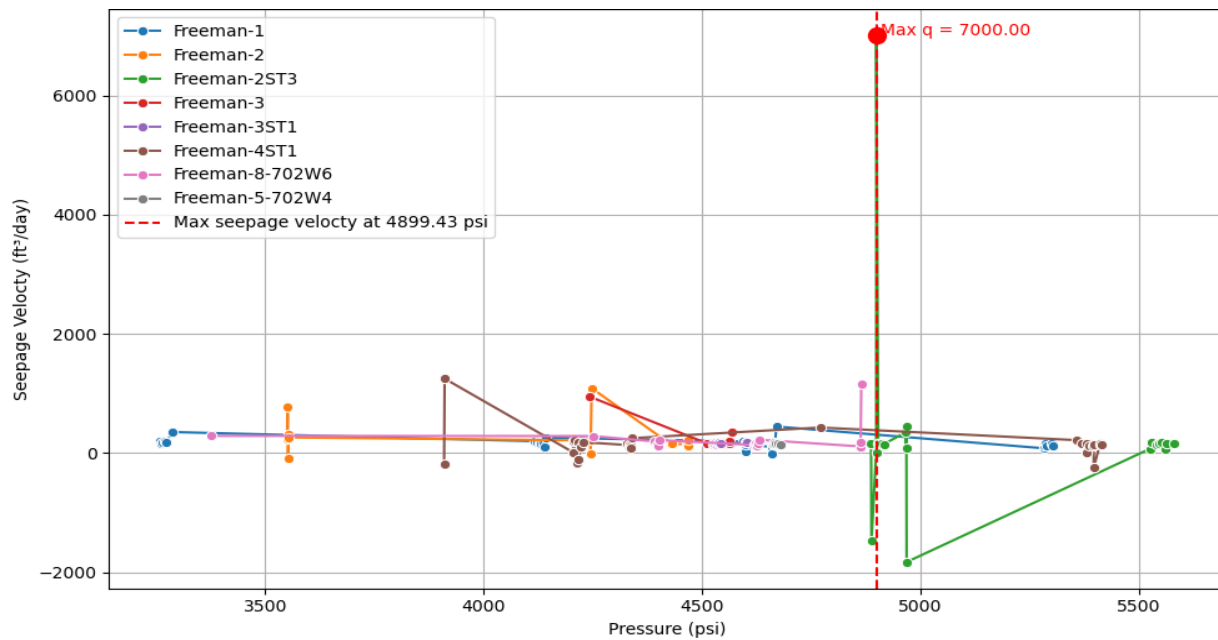


Figure 5a. The plot between seepage velocity versus pressure, showing a maximum seepage velocity of 7000 ft/day for well Freeman-2ST3 at a pressure of 4899 psi.

Characterization of fluid flow

The characterization of fluid flow dynamics across the study area reveals distinct pressure dependent regimes and a pivotal threshold range governing overall system behavior. Analysis of mean fluid flux, derived from bin averaged pressure data across all wells (Figure 5a), demonstrates an initial increase from approximately 3300 psi to a peak of 4000 psi, indicative of optimal bulk fluid transmissibility within this pressure window. Beyond this peak, flux values exhibit a sharp decline, stabilizing in positive territory until approximately 4800 psi. The pressure interval between 4800 psi and 5050 psi, represents a critical transition zone. Within this range, the mean fluid flux undergoes a dramatic shift from positive to negative values, coupled with an exceptionally large standard deviation, signifying a highly heterogeneous and unstable flow regime. This pronounced variability suggests the activation of complex subsurface processes, potentially including capillary breakthrough of sealing formations or the initiation of new fluid pathways (Espinoza and Santamarina, 2017).

Detailed investigation into individual well performance (Figure 5b), corroborates this interpretation, revealing a maximum seepage velocity of 7000 ft/day for well Freeman-2ST3 at a pressure of 4899 psi. The association of this high magnitude fluid event with oil (“o”) flow, precisely within the identified critical threshold, strongly indicates the activation of a highly conductive pathway or the breakthrough of a significant hydrocarbon (oil) column at

this specific pressure point. On the other hand, other wells within the same pressure range, exhibit widely disparate seepage velocities and fluid types (“g” for gas, “o/w” for oil/water), including instances of near-zero or negative flow. This heterogeneity underscores the localized nature of fluid transport and the significant influence of compartmentalization or specific geological features. Beyond 5050 psi, the mean fluid flux gradually stabilizes into a new, lower positive equilibrium regime (approximately 20 to 30 ft³/day), with reduced variability. These findings collectively establish the 4800 to 5050 psi pressure range as a critical threshold for this hydrocarbon system, where reservoir scale fluid flow dynamics are fundamentally altered, likely driven by exceeding precise petrophysical thresholds for fluid mobility and migration (Bai et al., 2021).

Determination of threshold pressure range (P_{range})

The threshold pressure range for highest fluid flux across all wells (Figure 4) and the threshold pressure range for seepage velocity across all wells (Figure 6), reveal critical insights into the pressure-dependent fluid flow dynamics of the reservoir system. Both plots, which are fundamentally related, measuring fluid transport through porous media, exhibit remarkably consistent trends concerning mean values and their associated standard deviations across varying pressure bin centres. This consistency underscores the robust identification of key flow regimes and a pivotal pressure threshold. Maximum pressure shown on the graph is approximately 5600 psi. This is the highest-pressure bin

centre displayed on the x-axis. From approximately 3300 psi to 5600 psi, both mean fluid flux and mean seepage velocity, demonstrate a positive increasing trend, culminating in a peak around 5600 psi. This initial enhancement in fluid transmissibility suggests that increasing reservoir pressure within this range is conducive to general fluid movement, potentially facilitating the initial stages of hydrocarbon migration by overcoming minor flow resistances. Following

this peak, both mean values sharply decline and then stabilize at relatively low positive levels until about 4800 psi, indicating a less efficient or more restricted bulk fluid transport regime in this intermediate pressure window. Both figures highlight a highly significant and common pressure range, approximately 4800 psi to 5600 psi, as a critical threshold (Zhang et al., 2022; Shagapov et al., 2023).

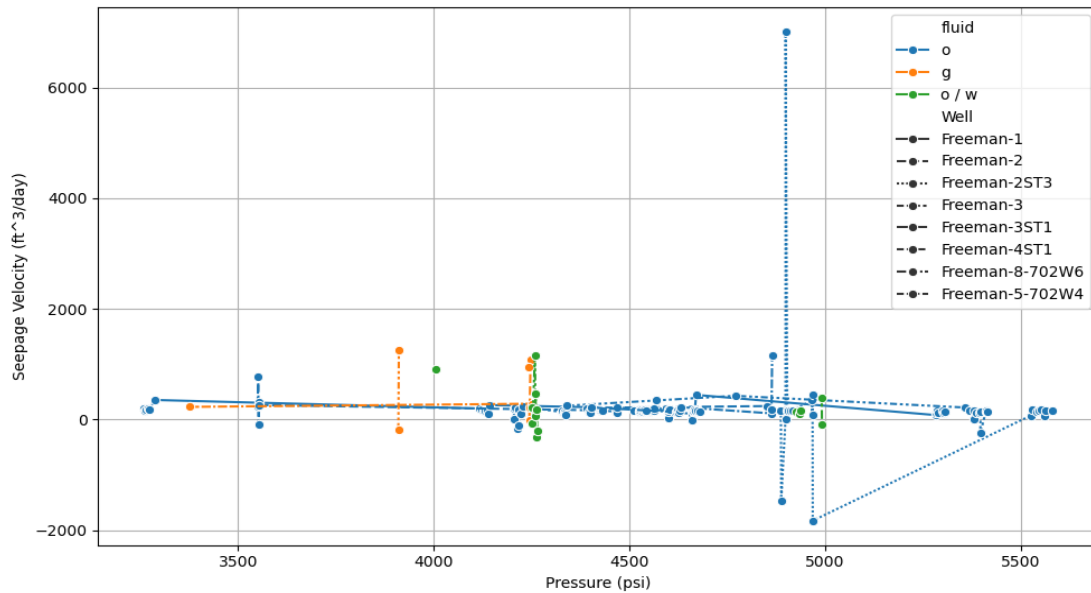


Figure 5b. The plot between seepage velocity and pressure revealing a maximum seepage velocity of 7000 ft/day for well Freeman-2ST3 at a pressure of 4899 psi including the type of fluid across the wells.

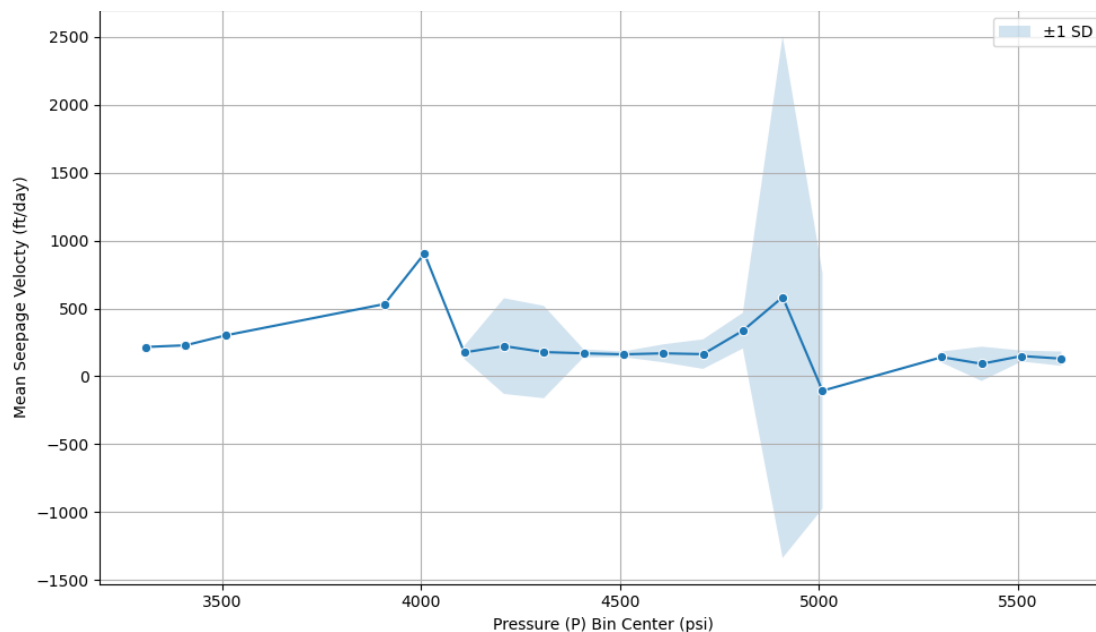


Figure 6. The plot between mean seepage velocity and pressure, showing threshold pressure range of 3300 psi to 5600 psi for seepage velocity of 900 ft/day.

Projection of fluid phases present across the pressure range

The fluid phases present across the pressure range were displayed in Figure 7. The plot primarily shows a high density of data points at fluid type code 1 (oil) throughout the entire observed pressure range (3300 psi to 5600 psi), indicating that oil is the predominant fluid encountered in the wells. Isolated occurrences of fluid type code 0 (gas), are observed at various pressures, including around 3400 psi, 3900 psi, 4200 psi, and near the 4800 to 5050 psi threshold. Moreover, instances of fluid type code 0.5 (oil/water) are present, notably around 4000 psi, 4200 psi, and within the 4800 to 5050 psi range. These figures provided a comprehensive understanding of the reservoir's behaviour. The extreme variability and directional shift in mean fluid flux observed in Figure 6 within the 4800 to 5050 psi pressure range, combined with the presence of oil, gas, and oil/water mixtures in Figure 7. within the same interval, strongly suggest that this represents a critical threshold for hydrocarbon mobility. This pressure range signifies the capillary breakthrough pressure of a sealing formation or the activation of high conductivity pathways, such as fractures or faults, which facilitate significant hydrocarbon migration (Rezaeyan et al., 2015; Zhang and Wang, 2022; Cheng et al., 2024).

The predominance of oil (fluid type code 1) throughout the entire pressure range from roughly 3,300 psi to 5,600 psi, suggests that the Freeman Field reservoirs are primarily oil-

charged, which is consistent with the Agbada Formation's well-established position as the Niger Delta's principal oil-producing interval. With a stable hydrocarbon column maintained throughout all recorded depth intervals, this oil dominance is indicative of a mature charge history. The isolated gas occurrences (code 0) at pressures of approximately 3,400, 3,900, and 4,200 psi, are explained as either dissolved gas that has separated from solution as local pressure falls below the bubble point or localized gas caps within structurally raised reservoir compartments. The presence of gas in the crucial 4,800–5,050 psi threshold range is especially important because it could be an indication of gas liberation or pressure-induced phase segregation related to the capillary breakthrough event detected at this period. In the studied wells, the oil/water mixture occurrences (code 0.5) around 4,000 and 4,200 psi indicate proximity to the oil-water contact (OWC), where log-derived fluid classification captures the transition zone between the hydrocarbon column and the aquifer. The reappearance of oil/water mixtures in the critical range of 4,800–5,050 psi, indicates dynamic upward movement of the OWC caused by pressure transients, which may indicate active water influx into the lower reservoir intervals when pressure drops. Taken together, Figure 7 show of spatial distribution of fluid phases supports the interpretation that the 4,800-5,050-psi pressure interval represents a zone of complex fluid-rock interaction, where simultaneous oil migration, gas exsolution, and water contact movement converge to produce the high flux variability seen in companion figures.

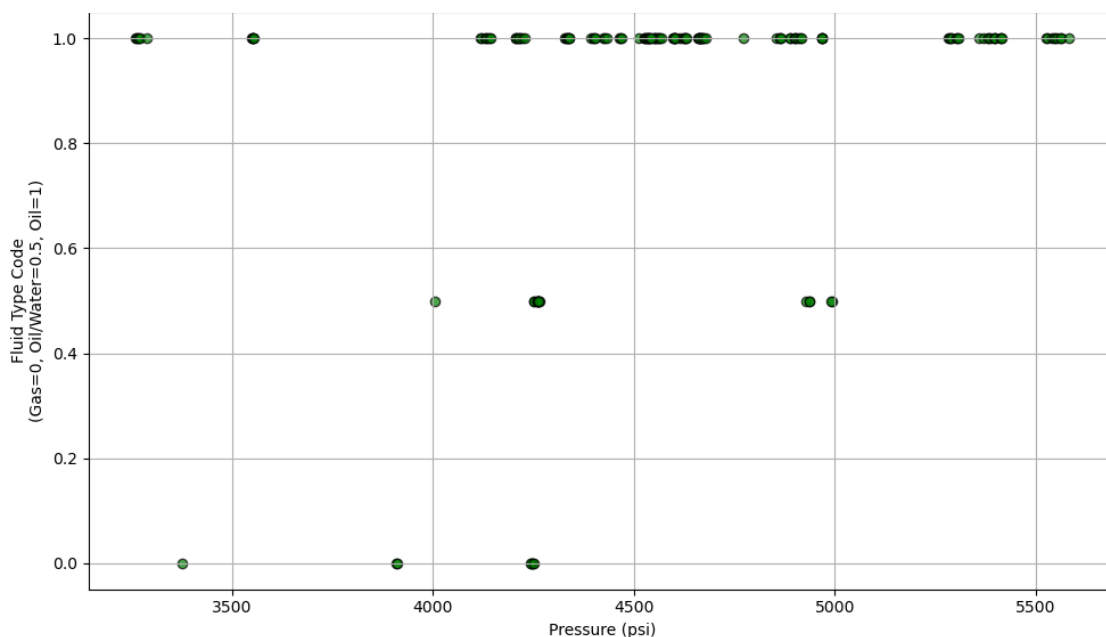


Figure 7. The plot of fluid type code versus pressure, showing the fluid phases present across the pressure range.

The occurrence of different fluid types within this highly dynamic pressure window further points to complex fluid-rock interactions or the movement of fluid contacts (oil-water contact, gas-oil contact) that dramatically impact bulk fluid flow. These findings are crucial for understanding the mechanisms of hydrocarbon accumulation and for optimizing future exploration and production strategies (Zhu et al., 2015; Li et al., 2023; Wang et al., 2025). Although capillary pressure curves were not available, the observed threshold behaviour is consistent with capillary-controlled flow mechanisms. Quantitative capillary pressure analysis (e.g., J-function) is recommended for future studies. Due to data confidentiality constraints, historical production data were not available for validation. However, the observed pressure–flux relationships are consistent with established reservoir flow behaviour.

CONCLUSIONS

This study used static pressure and wireline log data from eight Freeman Field wells, to examine pressure-dependent fluid flow dynamics in a Niger Delta reservoir. The fluid flux, seepage velocity, and pressure thresholds controlling hydrocarbon migration were measured using modified Darcy's Law. Some of the key findings reached as follows:

1. The lowest pressure needed to start a major hydrocarbon migration in the study reservoir was determined to be a threshold pressure (P_{thrsh}) of roughly 4899 psi. Fluid flux stays low at this pressure, indicating that there is not enough pushing energy to get past capillary barriers in the pore network.
2. The Freeman Field reservoir system has a maximum fluid flux of 1400 ft³/day at approximately 4899 psi (P_{max}), indicating peak hydrocarbon mobility.
3. The pressure range of 4800 and 5050 psi has been recognized as the crucial ideal window for continuous hydrocarbon flow. High flux variability and the presence of oil, gas, and oil/water phases within this interval are interpreted as signs of capillary breakthrough or the activation of high-conductivity fracture channels.
4. In every sampled pressure range (4800 – 5050 psi), oil is the predominant fluid phase. Gas and oil/water mixtures are defined as proximity to the oil-water contact and localized gas caps or exsolved gas, respectively, and occur at discrete pressure intervals.
5. Maintaining reservoir pressure over the defined threshold of 4899 psi is critical to ensuring hydrocarbon mobility and production efficiency. The quantitative pressure windows obtained here serve as a direct foundation for the planning of pressure-maintenance EOR activities,

such as waterflooding and gas injection, in the study area and similar Niger Delta reservoirs.

The Python-based multi-well computational approach that combines Darcy's Law with petrophysical log and static pressure data, provides a repeatable, cost-effective framework for pressure migration analysis that may be applied to other heterogeneous basins.

ACKNOWLEDGMENTS

The authors thank the Nigerian Upstream Petroleum Regulatory Commission, Lagos, Nigeria, and Shell Petroleum Development Company, SPDC, Nigeria, for emancipating the data from which this study was conducted.

Author Credit Statement

Olusegun Alabi and Abigail Olaoluwa developed the concept and design this work. Enock Oladimeji and Bosede Ojo did the analysis and interpretation of the data while Samuel Sedara and Abigail Olaoluwa did the writing and editing of the manuscript. In general, Olusegun Alabi supervised and validated the work. The authors approved the submission and publication of the manuscript and therefore take public responsibility for the research

Data Availability

The data that support the findings of this study are available from the corresponding author upon request.

Compliance with Ethical Standards

The authors declare no conflict of interest and adhere to copyright norms

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Received on: 12-09-2025; Revised on: 1-7-2026; Accepted on: 2-7-2026

Petrology and mineral chemistry of Mesoproterozoic granitoids from the southwestern Chhotanagpur Granite Gneiss Complex (CGGC), Central India: Physico-chemical conditions of crystallization

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ABSTRACT

We present field relationships, petrographic features, and mineral chemistry of feldspar, biotite, and amphibole for the Mesoproterozoic granitoids from the southwestern part of the Chhotanagpur Granite Gneiss Complex (CGGC) in central India. Based on field and petrographic observations, the granitoids in the study area are characterized as porphyritic granite and equigranular granite. The P-T- fO_2 conditions and physico-chemical environment during the emplacement of these granitoids have been established based on mineral compositions of amphibole and biotite. The estimated temperatures from amphibole and biotite thermometry are similar with ranges of 738-769 °C for porphyritic granite and 723-798 °C for equigranular granite, respectively. The calculated barometry based on amphibole and biotite shows pressures ranging from 3.2 to 4 kbar for porphyritic granite, from 2 to 2.8 kbar for equigranular granite, and attaining up to 5.7 kbar for the equigranular granite. The estimated barometric conditions suggest the crystallization of porphyritic granite at a depth of ~13 km, and the equigranular granite, at ~14 km, corresponding to the upper-crustal crystallization conditions. The amphibole composition indicates that the granite crystallized from a volatile-rich magma under high oxygen fugacity conditions. Biotite and amphibole thermobarometric conditions of equigranular granite, indicate that these rocks crystallized under a high-oxidizing environment buffered in NNO (nickel-nickel-oxygen) to HM (hematite-magnetite) conditions. Biotite compositions indicate these granitoids to have been derived from a crustal source due to alkaline A-type parental magma in an anorogenic tectonic environment.

Keywords: Mineral chemistry; Thermobarometry; Mesoproterozoic; A-type granitoids; Chhotanagpur Granite Gneiss Complex (CGGC); oxygen fugacity.

INTRODUCTION

Granitoids constitute a significant part of the continental crust and exhibit diverse chemical compositions. Petrogenetic studies of granitoids, play a vital role in unravelling the crustal evolution and crust-mantle interaction across different tectonic settings, providing a broader picture of crustal growth (Campbell and Taylor, 1983; Annen et al., 2006; Cawood et al., 2022). Understanding the geochemical diversity in granitoids of calc-alkaline and alkaline affinity under different temperature (T), pressure (P), and oxygen fugacity (fO_2) conditions, helps to decipher the physico-chemical conditions of magma crystallization and the tectono-magmatic evolution (Wones and Eugster, 1965; Speer, 1987; Vyhnal et al., 1991; Abdel-Rahman, 1994; Ague, 1997; Bhattacharya et al., 2014).

Amphibole and biotite are two common ferromagnesian mineral phases present in granitic rocks, which are sensitive to changes in pressure, temperature, and oxygen fugacity (fO_2) (Putirka, 2016; Joshi et al., 2021; Gion et al., 2022; Bartoli et al., 2024). Thus, compositions of these mineral phases are utilized to determine the thermobarometric conditions during the emplacement of granitic rocks. The compositions of these minerals are substantially influenced by the magma chamber processes, their source characteristics, and tectonic setting (Abdel-Rahman, 1994; Elliott, 2001; Bachmann and Dungan, 2002; Kumar and Pathak, 2010; Ridolfi et al., 2010; Elangovan et al., 2020). Studies have shown that aluminium and titanium contents in amphibole are sensitive to crystallization depth, temperature, pressure, and oxygen fugacity (fO_2) (Schmidt,

1992; Ridolfi et al., 2010). Several parameters, such as the Al-in-hornblende (Hammarstrom and Zen, 1986; Schmidt, 1992; Ridolfi et al., 2010), Ti-in-hornblende (Ottens, 1984), coexisting hornblende-plagioclase compositions (Holland and Blundy, 1994), and water contents in amphibole (Naney, 1983; Dall'Agnol et al., 1999; Parat et al., 2008; de Oliveira et al., 2010), are widely used to measure the geothermobarometric conditions and stability depth for amphibole crystallization. Similarly, biotite is also sensitive to the crystallization conditions of magma, such as pressure, temperature, fO_2 , and water contents (Wones and Eugster, 1965; Czamanske and Wones, 1973). It has been demonstrated that the chemistry of biotite provides reliable constraints on parameters such as oxybarometer, melt composition, and the tectonic setting (Czamanske and Wones, 1973; Abdel-Rahman, 1994; Shabani et al., 2003). Redox conditions of granitic magma (Kumar and Pathak, 2010; Raju et al., 2022) correspond to the biotite end-members, such as annite (Fe^{2+}) and oxynite (Fe^{3+}). These end-members of biotite are also associated with the magnetite- and ilmenite-series of Ishihara (1977), respectively.

The Chhotanagpur Granite Gneiss Complex (CGGC) (Figure 1a) lies in central India and represents an extended crustal segment of the Proterozoic Central Indian Tectonic Zone (CITZ) (Acharyya, 2003; Sanyal and Sengupta, 2012). The basement complex of the CGGC is intruded by several episodes of calc-alkaline and alkaline granitic magmatism during the Proterozoic. Nonetheless, the granitoids are the major components of the CGGC and exhibit a wide range of

mineralogical variation. Recent studies, based on U-Pb zircon geochronology, bulk rock geochemistry, and isotopic compositions have attempted to construe the nature of protolith and tectonic setting of the different phases of granitoids in the CGGC (Saikia et al., 2017; Mukherjee et al., 2018, 2019; Chakraborty et al., 2019; Sequeira et al., 2022). The older phases of granitoids and migmatites of ca. 1660 Ma have been reported to be formed in an arc environment (Mukherjee et al., 2017; Saikia et al., 2017), ca. 1450 Ma porphyritic granites in an extensional setting (Mukherjee et al., 2018; Sequeira et al., 2022), and the ca. 1140 Ma equigranular granites emplaced in a post-collisional tectonic regime (Qasim et al., 2025).

However, the available information on the petrogenesis of CGGC granitoids and their tectonic discrimination based on mineral chemistry and the P-T- fO_2 conditions of emplacement is poorly constrained. Additionally, the source magma of these granitoids, whether they were derived from the crust or crust-mantle mixed source remains uncertain. Nonetheless, a few studies have focused on mineral chemistry of granitoids to constrain the source and tectonic setting(s) of the CGGC granitoids (Ghose, 1983; Goswami and Bhattacharya, 2014; Goswami et al., 2018; Saikia et al., 2019; Das et al., 2020; Roy et al., 2020), yet they are less understood in terms of mineral

characterization and physico-chemical conditions of granitic magma. In the present study, we present field relationships, petrographic, and mineral chemistry data for two Mesoproterozoic granitoids from the southwestern part of the CGGC. The objectives of the present study are to characterize the minerals, to constrain the crystallization conditions during the emplacement of these granitoids, and to understand the tectonic evolution of the region.

GEOLOGICAL BACKGROUND

The Chhotanagpur Granite Gneiss Complex (CGGC) in central India (Figure 1a) covers an area of ~100,000 km² and is recognized as the eastern extension of the CITZ. The terrane is demarcated by the Gangetic sediments to the north, North Singhbhum Fold Belt to the south, Rajmahal Traps and Bengal Basin sediments to the east, and Vindhyan Supergroup to the northwest (Figure 1a). The western part of the CGGC is majorly covered by Gondwana Supergroup sediments and Deccan volcanics (Figure 1a). The rocks of CGGC are characterized as granitoid gneisses, migmatites, granites, calc-silicates, patches of metapelites, mafic intrusive bodies and dykes, pegmatites, and aplitic veins (Ghose, 1983; Acharyya, 2003; Sanyal and Sengupta, 2012; Patel et al., 2021).

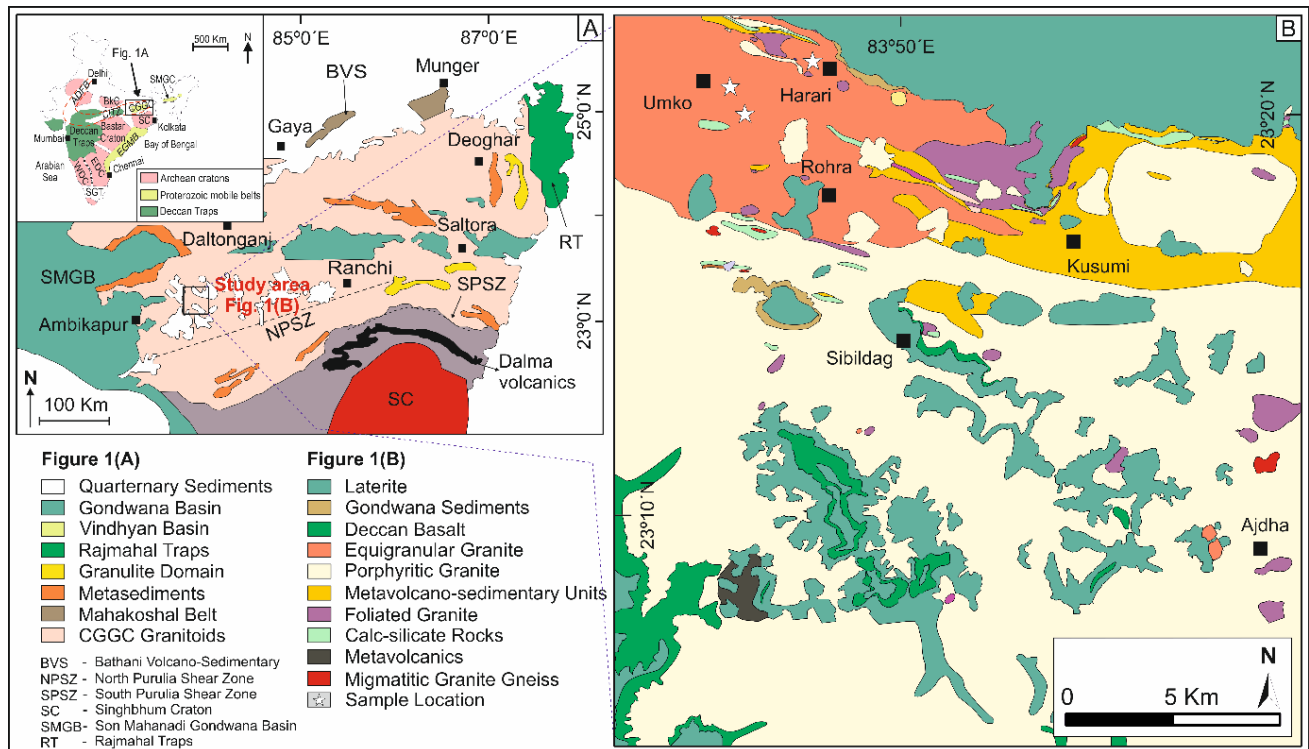


Figure 1. Inset map of peninsular India showing major Archean tectonic divisions and Proterozoic orogenic belts with tracts of Central Indian Tectonic Zone (CITZ); (a) Simplified geological map of Chhotanagpur Granite Gneiss Complex (CGGC) showing the distribution of major lithological units (Acharyya, 2003), and the study area is indicated by a black rectangle; (b) Simplified geological map of the study area showing the major lithological units (modified after Mukherjee and Kaushik, 2019; Ghosh and Patel, 2021; Qasim et al., 2025).

Geochronological studies have shown that the emplacement of porphyritic granitoids occurred between c. 1750 and 1660 Ma in the western and northern CGGC (Chatterjee and Ghose, 2011; Saikia et al., 2017). The emplacement of massif anorthosite at c. 1550 Ma was documented from the southeastern part (Chatterjee et al., 2008), followed by Mesoproterozoic (ca. 1450 Ma) ferroan A-type granitoids from the eastern parts (Mukherjee et al., 2018; Sequeira et al., 2022). The emplacement of ca. 1140–1005 Ma granitoids from the southwestern CGGC (Singh and Krishna, 2009; Qasim et al., 2025) correlated with the final phase of the continent-continent collision of northern and southern Indian blocks (Bhowmik et al., 2011).

The CGGC has experienced multiple episodes of magmatism, varying degrees of metamorphism, and tectonic deformation, as reported across different segments of the terrane (Maji et al., 2008; Rekha et al., 2011; Sanyal and Sengupta, 2012; Saikia et al., 2017). Available age data suggest four phases of metamorphic imprints in the CGGC (Ghose, 1983). The metamorphic event M1 occurred during ca. 1680–1580 Ma, corresponding to the ultra-high temperature metamorphism, resulting in extensive dehydration melting, felsic magmatism, and later the emplacement of Saltora gabbroic anorthosite at ca. 1550 Ma (Chatterjee et al., 2008, 2010; Rekha et al., 2011; Sanyal and Sengupta, 2012; Dey et al., 2019; Mukherjee et al., 2019). The metamorphic event M2 corresponds to the migmatization of ca. 1660 Ma granitoids, followed by the emplacement of ca. 1450 Ma granitoids (Sanyal and Sengupta, 2012; Mukherjee et al., 2018; Kumar and Dwivedi, 2019; Sequeira et al., 2022). The granulite facies metamorphic event (M3) associated with the Grenvillian orogeny occurred during ca. 1100–950 Ma, and the continent-continent collision between northern and southern Indian blocks at ca. 1000 Ma (Acharyya, 2003; Maji et al., 2008; Karmakar et al., 2011; Rekha et al., 2011; Dey et al., 2019). The last major

metamorphic event (M4) occurred at ca. 870–780 Ma, culminating at peak P-T conditions of ~750–600 °C at 7–12 kbar (Chatterjee et al., 2010; Ray et al., 2011; Sanyal and Sengupta, 2012).

FIELD FEATURES AND PETROGRAPHY

The study area lies in the east of Ambikapur town near Umko and Harari villages, constituting the marginal segment of the southwestern CGGC (Figure 1a). Major lithologies of the study area are granitic gneiss, granitoids, amphibolites, mafic and felsic intrusive rocks (Ghosh and Patel, 2021) (Figure 1b). Subordinate metavolcano-sedimentary supracrustal rocks occur in the study area with co-deformed interlayers of metapelite, quartzite, and calc-silicate rocks (Figure 1b). Mafic dykes, tourmaline-bearing pegmatites, quartz, and aplitic veins occur as intrusives. The Lower Gondwana sediments, Deccan Trap basalts, and aluminous laterite cover the rocks of the study area. Based on the field relationships, macroscopic features, and petrographic studies, granitoids in the study area have been divided into two types: porphyritic granite and equigranular granite. A summary of the sample details, locations, petrographic features, and mineral compositions is presented in Table 1.

Porphyritic granite

Porphyritic granites are the dominant lithologies exposed in the study area (Figure 1b). They are pink to grey, medium- to coarse-grained, and exhibit porphyritic texture. The mineralogical composition of porphyritic granite is dominated by quartz (25–28 %), microcline (20–25 %), plagioclase (18–20 %), orthoclase (12–14 %), and biotite (10–12 %) (Appendix 1). Titanite, zircon, and apatite are present as accessory minerals. Porphyritic granite exhibits characteristically large phenocrysts of K-feldspar surrounded by coarse-grained groundmass composed of quartz, K-feldspar, plagioclase, and biotite (Figure 2a).

Table 1. Description of samples, locations, petrographic features, and mineral compositions of granitoids from the southwestern part of CGGC

Sr. No.	Sample No.	Rock type	Location	Latitude	Longitude	Major minerals	Accessory minerals	Deformation	Elemental compositions (wt.%)		Ratio	
									K-feldspar	Plagioclase	Biotite	Amphibole
1	CG-1	Porphyritic granite	Harari	23°21'21" N	83°47'17" E	Qz+Afs+Pl+Mc+Bt+Myr	Apt+Ser+Opq+Zrn	Undeformed	Or = 90 - 94 Ab = 6 - 10	Ab = 86 - 96 An = 9 - 14	X _{Mg} = 0.23 - 0.27	
2	CG-2	Equigranular granite	Umko	23°19'37" N	83°44'09" E	Qz+Afs+Pl+Mc+Bt+Amp	Ttn+Ser	Undeformed	Or = 94 - 95 Ab = 5 - 7	Ab = 55 - 82 An = 17 - 44	X _{Mg} = 0.37 - 0.41	Mg# = 0.51 - 0.56
3	CG-4	Equigranular granite	Umko	23°19'37" N	83°45'03" E	Qz+Afs+Pl+Mc+Bt	Ttn+Ser+Opq	Undeformed	Or = 94 - 95 Ab = 5 - 7	Ab = 55 - 82 An = 17 - 44	X _{Mg} = 0.37 - 0.42	

Abbreviations: Qz = Quartz, Pl = Plagioclase, Afs = Alkali Feldspar, Mc = Microcline, Bt = Biotite, Amp = Amphibole, Ser = Sericite, Ttn = Titanite, Opq = Opaques, Zrn = Zircon, Apt = Apatite, Myr = Myrmekite

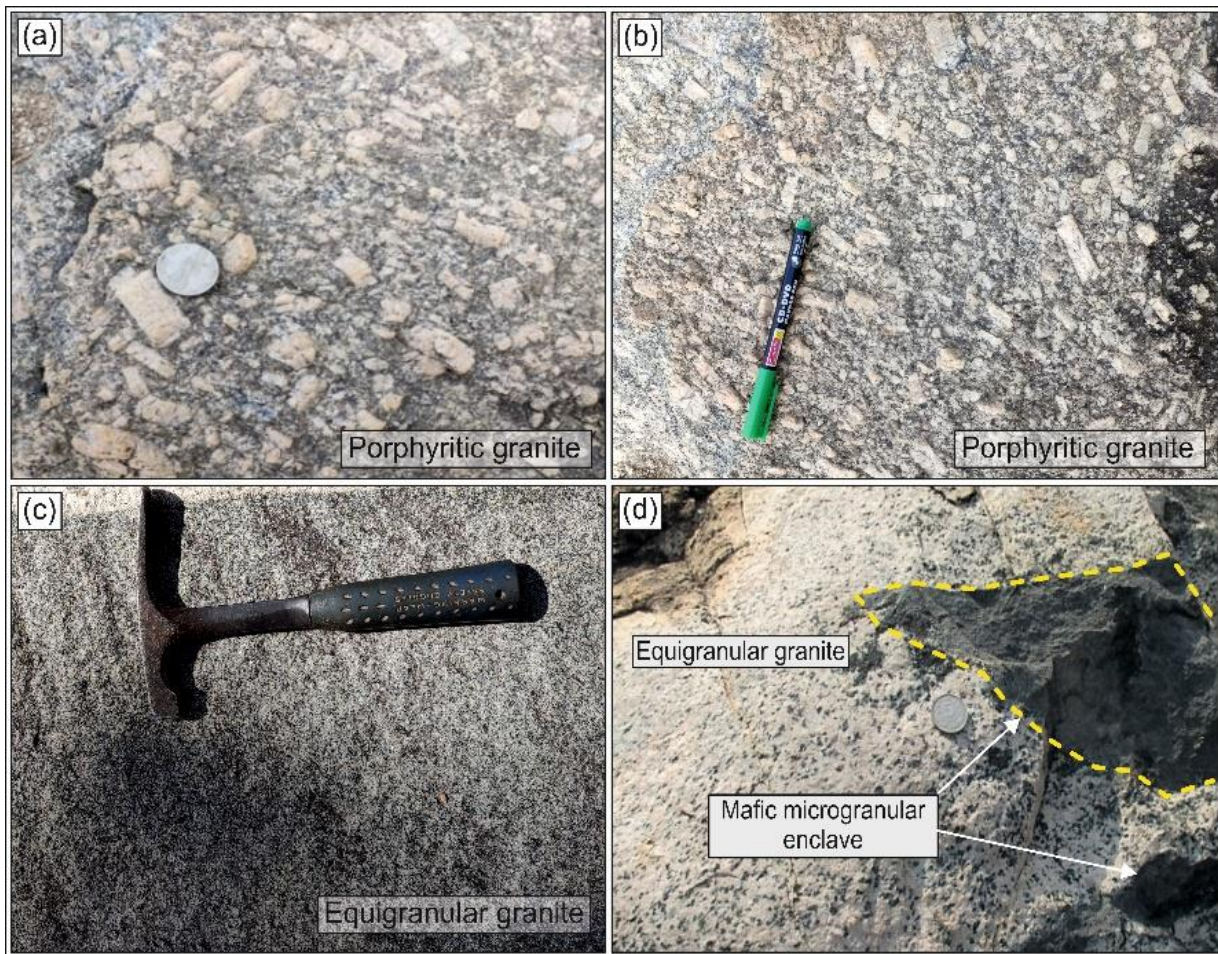


Figure 2. Representative field photographs from the southwestern part of CGGC. (a, b) Porphyritic granite exhibiting the dominant K-feldspar megacrysts up to ~2 cm in size enclosed by quartz and biotite groundmass near Harari; (c) Undeformed equigranular granite showing medium-grained uniform texture near Umko; (d) Mafic microgranular enclave (MME) hosted by equigranular granite.

The K-feldspar phenocrysts show predominantly random orientation, locally exhibiting crude alignment of longer axes (Figure 2b). To some extent, the induced recrystallization affected the rock, resulting in sub-rounded to elliptical K-feldspar megacrysts, reabsorbed quartz and feldspar, and sutured grain boundaries (Figure 3a). These rocks have preserved the primary magmatic features. Porphyritic granites are composed of quartz, alkali feldspar, plagioclase, and biotite with accessory minerals of titanite, zircon, and secondary muscovite (Figure 3a). Biotite occurs as the major ferromagnesian phase. Microcline shows cross-hatched twinning. Sericitization of plagioclase and perthite exsolution in orthoclase is observed in this rock.

Equigranular granite

The equigranular granite samples were collected from the Umko area (Figure 1b). They are coarse to medium-grained, equigranular, undeformed, leucocratic, grey to pink-coloured, and occur as small plug-like bodies in the study area (Figure 2c). The mineralogical composition of equigranular granite is

dominated by quartz (20–25 %), microcline (10–20 %), plagioclase (16–22 %), orthoclase (12–18 %), biotite (12–18 %), and amphibole (8–10 %) (Appendix 1). Zircon, titanite, and apatite occur as accessory minerals. In the field, the equigranular granites are undeformed and exhibit a homogeneous texture, reflecting minimal overprinting by later metamorphism and deformation (Figure 2c). Mafic microgranular enclaves (MME) of varying shapes and sizes are observed in this granite (Figure 2d). The MMEs are fine-grained, made of amphibole, biotite, plagioclase, and minor quartz. The equigranular granites are primarily composed of quartz, alkali feldspar, and plagioclase with a hypidiomorphic texture (Figure 3b-d). Biotite and hornblende are the ferromagnesian mineral phases (Figure 3e). The back-scattered electron (BSE) image of equigranular granite shows that the biotite is euhedral and devoid of hydrothermal alteration or chloritization (Figure 3f). Other accessory mineral phases include titanite, zircon, apatite, and opaques. Perthitic unmixing of alkali feldspar and sericitization of plagioclase are common in these rocks. An increase in the modal proportion of mafic minerals is observed near MME.

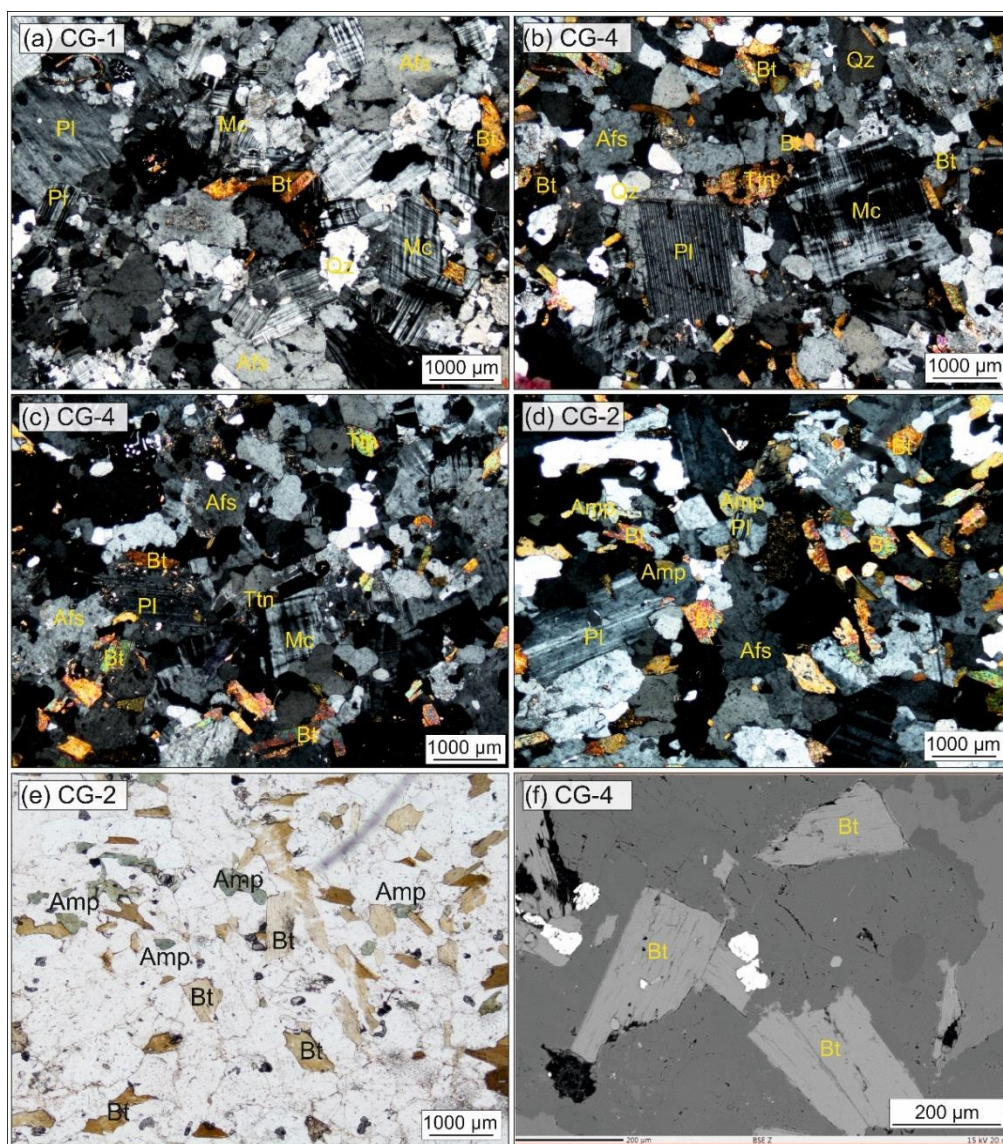


Figure 3. Representative photomicrographs of the studied granitoids samples. (a) Porphyritic granite is majorly comprising quartz, alkali feldspars, plagioclase, and biotite; (b-d) Equigranular granite samples are displaying the hypidiomorphic textures with association of ferromagnesian mineral assemblages of biotite and amphibole with accessory titanite; (e) Equigranular granite sample (captured in plane-polarized light) is showing the mineral assemblages of biotite and amphibole; (f) Back-scattered electron (BSE) image of equigranular granite shows the magmatic biotite is devoid of hydrothermal alterations. (Abbreviations: Qz-Quartz, Afs-Alkali-feldspar, Pl-Plagioclase, Bt-Biotite, Amp-Amphibole, Mc-Microcline, and Ttn-Titanite).

ANALYTICAL TECHNIQUES

Three representative granitoid samples (one porphyritic granite and two equigranular granites) with the least alteration, were selected for mineral composition analysis of biotite, amphibole, and feldspar using an electron probe microanalyzer (EPMA). For the microprobe analysis, thin sections were well polished with diamond paste (Metatech, 1 µm and ¼ µm), and a 20 nm carbon coating was applied using the LEICA-EM ACE 200 carbon coater. The mineral compositions were analyzed by CAMECA SX Five electron microprobe under the condition of an accelerating voltage of 15 kV, a 10 nA beam current with a LaB₆ filament source in the electron gun for producing an

electron beam of 1 µm diameter at the Department of Geology, Banaras Hindu University (BHU), India. The analyses were carried out using wavelength dispersive spectrometry with TAP, LIF, LPET, LTAP, and PET crystals. The standards used for calibrations are, GeoStd/Ab (albite), GeoStd/Or (orthoclase), GeoStd/Prdt (peridotite), GeoStd/Di (diopside), and GeoStd/Alm (almandine). The SxSAB version 6.1 and the CAMECA software SX-Results were used for calibration, acquisition, quantification, and data processing. Based on repeated standard calibration analyses (Pandey et al., 2017), the precision of the analysis was estimated to be better than 1% for major-element oxides.

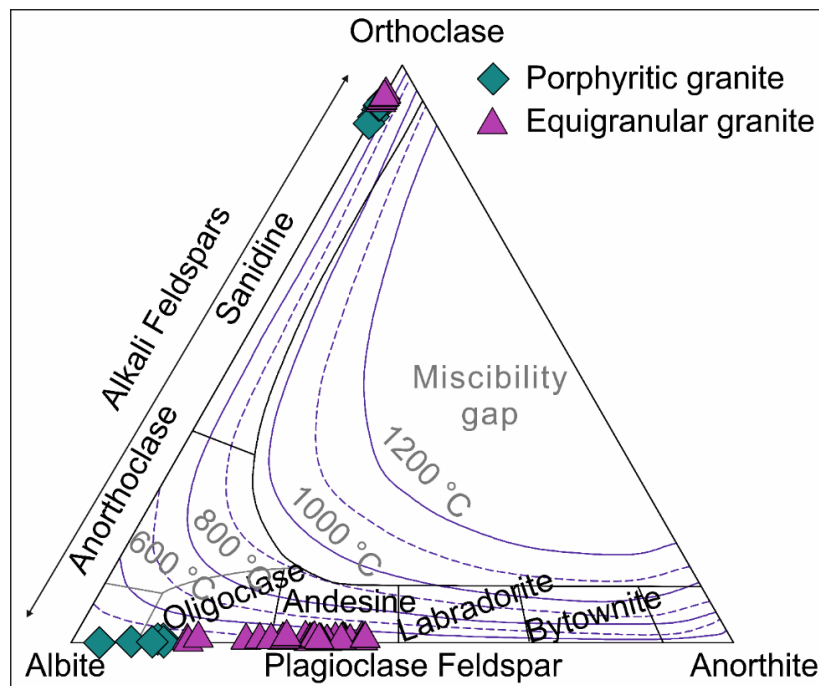


Figure 4. Ab-Or-An ternary feldspar diagram (Deer et al., 1992) showing the compositional variations of feldspars from the studied granitoids. The isotherms curves for temperature are shown based on the proposed model after Elkins and Grove (1990).

The structural formula of biotite analyses was calculated based on 22 oxygen atoms, following Li et al. (2020), and stoichiometric charge-balance calculations estimated the partitioning of Fe^{2+} and Fe^{3+} from total FeO during cation normalization using the MagMin_PT Excel Programme (Gündüz and Asan, 2023).

RESULTS

Mineral chemistry

Alkali-feldspar

A total of 19 point analyses on alkali feldspars (12 points on porphyritic granite and 7 points on equigranular granite), were obtained from the studied granitoids. The compositions of alkali feldspar obtained from the microprobe analysis are presented in Appendix 2. The analyzed alkali feldspar results are illustrated on the Ab-Or-An ternary diagram (Figure 4). The compositions of alkali feldspar range from Ab_{5-7} to Or_{94-95} in equigranular granite and from Ab_{6-10} to Or_{90-94} in porphyritic granite.

Plagioclase

A total of 37 point analyses on plagioclase (6 points on porphyritic granite and 31 points on equigranular granite) were obtained, and the results are given in Appendix 3. Plagioclase compositions of granitoids are illustrated on the Ab-Or-An ternary diagram (Figure 4). They range in composition from oligoclase to andesine (Ab_{82-55} to An_{44-17}) in equigranular

granite and are confined to oligoclase (Ab_{96-86} to An_{14-9}) in porphyritic granite.

Amphibole

Amphiboles occur exclusively in equigranular granite. Accordingly, analyses of 7 points from equigranular granite were obtained, and the results are presented in Appendix 4. The structural formula of amphibole was calculated based on 23 oxygen atoms following the International Mineralogical Association (IMA) scheme (Hawthorne et al., 2012). In amphibole, Fe^{3+} was estimated stoichiometrically using the internal normalization and site-allocation procedure implemented in the Ridolfi (2021) amphibole thermos barometry. Amphibole compositions of equigranular granite have high total alkalis (Na+K) with a majority of >0.1 Ti atoms per formula unit (apfu) (Czamanske and Wones, 1973; Girardeau and Mevel, 1982; Keeditse et al., 2016), suggesting primary amphibole. Additionally, in the (Ca + Na + K) versus Si^{IV} diagram (Czamanske and Wones, 1973), all the amphiboles exhibit a magmatic nature. Equigranular granite amphiboles have high Si contents (>6.5 apfu) and Mg# values (0.34–0.37). Based on the amphibole classification of Leake et al. (1997), amphiboles in equigranular granite have ferroedenite affinity (Figure 5a). In amphibole composition diagrams (Figure 5b, c), equigranular granite amphiboles are distributed into the intermediate $f\text{O}_2$ field and confined to above the nickel-nickel-oxygen (NNO) buffer, respectively.

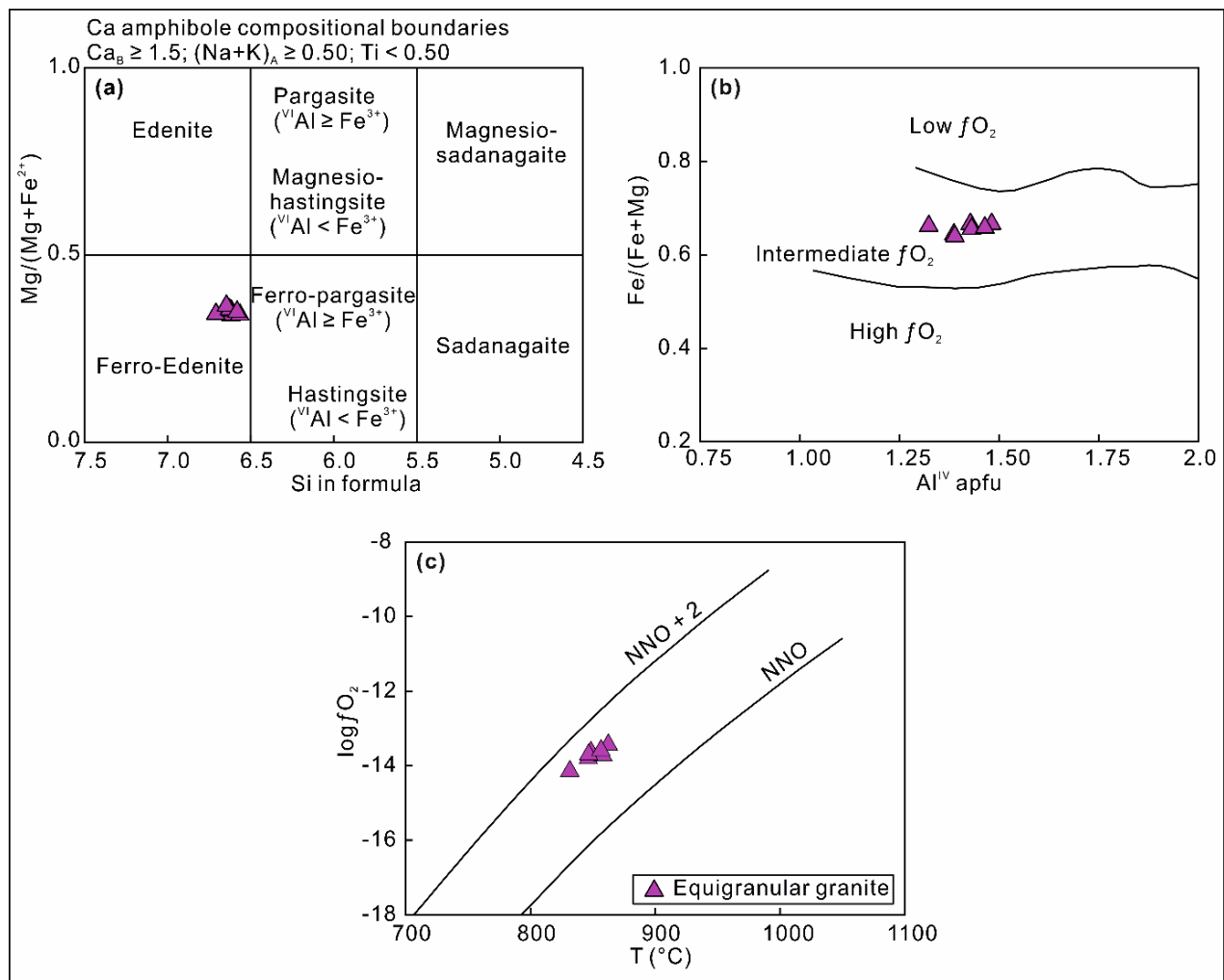


Figure 5. Amphibole classification diagrams. **(a)** Mg/(Mg+Fe²⁺) vs. Si diagram (Leake et al., 1997) showing the amphiboles of equigranular granite are calcic in nature and consistent with ferro-edenite affinity; **(b)** Fe/(Fe+Mg) vs. Al^{IV} diagram (Anderson and Smith, 1995) showing the moderate oxygen fugacity (fO_2) for equigranular granite; **(c)** Log fO_2 vs. T (°C) diagram (Ridolfi et al., 2010) indicating that the amphiboles from equigranular granite buffered at NNO (nickel-nickel-oxygen).

Biotite

A total of 32 point analyses on biotite (5 points on porphyritic granite and 27 points on equigranular granite), are obtained from the studied granitoids. The analyzed element compositions of biotite grains are presented in Appendix 5. The biotite structural formula was calculated using the method described by Li et al. (2020), based on 22 oxygen atoms. Stoichiometric charge-balance calculations estimated Fe³⁺ iron during the cation normalization. All the biotites of the studied granitoids have low Ti (<0.5 apfu), Al^{VI} (<1 apfu) contents, and

moderate to low Mg/(Mg+Fe) ratios (Appendix 5). The X_{Mg} { $X = Mg/(Mg+Fe)$ } values of biotites in porphyritic granite are comparatively lower (0.23–0.27) than those of biotites in equigranular granite ($X_{Mg} = 0.37$ –0.42) (Table 2). On the compositional triangular diagram (Figure 6a), biotites from the studied samples are found in the Fe-biotite fields with high iron contents. Likewise, on the ternary biotite classification diagram (Figure 6b), they straddle the boundary between primary biotite and re-equilibrated primary biotite fields. On the classification diagram (Figure 6c), all biotites show compositional variation between annite and siderophyllite.

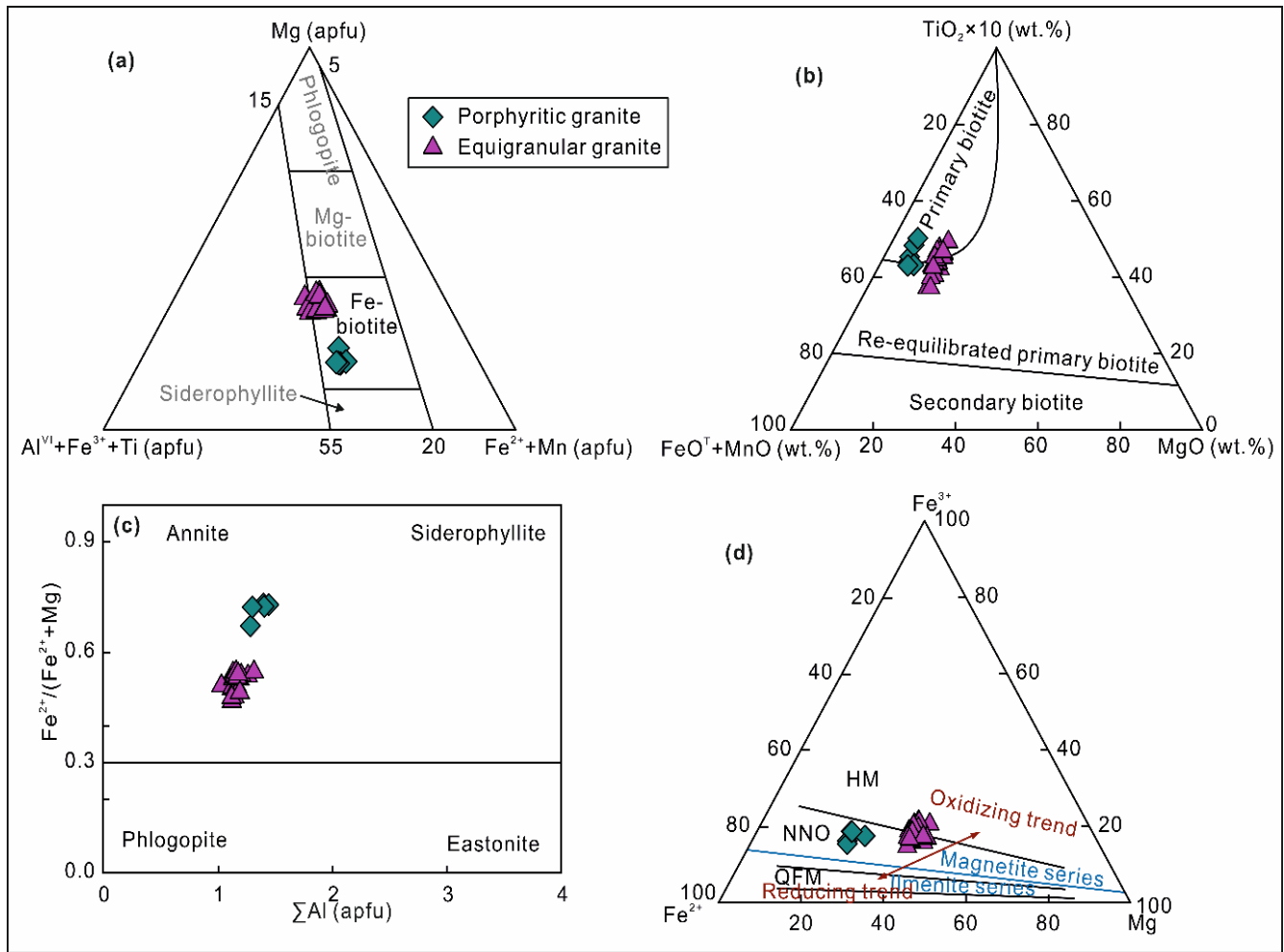


Figure 6. (a) $(Al^{VI}+Fe^{3+}+Ti)-Mg-(Fe^{2+}+Mn)$ ternary biotite classification diagram (Foster, 1960), showing the biotite from the studied granitoids are Fe-biotite; (b) $(FeO^T+MnO)-10\times TiO_2-MgO$ ternary biotite diagram (Nachit et al., 2005), showing the studied granitoids biotite are consistent with primary biotite and re-equilibrated primary biotite composition; (c) $Fe^{2+}/(Fe^{2+}+Mg)$ vs. ΣAl (apfu) biotite classification diagram (Rieder et al., 1998); (d) $Fe^{2+}-Fe^{3+}-Mg$ ternary biotite diagram (Wones and Eugster, 1965), indicating oxidizing nature with NNO-HM buffers for the studied granitoids. Abbreviations: HM = hematite-magnetite, NNO = nickel-nickel-oxygen, QFM = quartz-fayalite-magnetite.

DISCUSSION

Crystallization conditions

The compositions of biotite and amphibole are resilient to pressure (P), temperature (T), oxygen fugacity ($\log fO_2$) conditions, and water content (H_2O_{melt}) at which they equilibrate or crystallize (Foster, 1960; Wones and Eugster, 1965; Anderson and Smith, 1995; Zhang et al., 2015; Bucholz et al., 2018). These parameters can be used to evaluate the intensive variables during the crystallization of host rocks. In this study, we utilized the MagMin_PT Excel Programme (Gündüz and Asan, 2023) to recalculate amphibole and biotite chemistry following the methods of Ridolfi (2021) and Li et al.

(2020). The barometric conditions of amphibole and biotite were calculated with accepted uncertainties of ± 1 kbar, as provided by Ridolfi (2021), and ± 1.27 kbar, as provided by Li and Zhang (2022), respectively. The biotite-only thermobarometric model of Li and Zhang (2022) was used to calculate the thermobarometry conditions of biotite. While adapting the method by Li and Zhang (2022), we followed the suggested recommendations, such as excluding the analytical spots with $K_2O < 5$ wt.% and the sum of total cations at tetrahedral (T), octahedral (M), and interlayer (A) sites exceeding 10% of their corresponding maximum theoretical values. A summary of calculated thermobarometry conditions is presented in Table 2.

Table 2. Summary of thermobarometry conditions of amphibole and biotite of granitoids from the southwestern part of CGGC

Indicator	Rock type	Porphyritic granite		Equigranular granite	
Parameters	Method	Min	Max	Min	Max
Temperature (°C)	Amphibole (Ridolfi, 2021)			723	741
	Biotite-only thermobarometric method (Li and Zhang, 2022)	738	769	769	798
Pressure (kbar)	Amphibole (Ridolfi, 2021)			1.97	2.61
	Al-in-hornblende geobarometry (Anderson and Smith, 1995)			2.09	2.82
	Biotite-only thermobarometric method (Li and Zhang, 2022)	3.19	4.04	2.97	5.69
LogfO ₂	Amphibole (Ridolfi, 2021)			-15.6	-14.7
	Biotite (Wones, 1989)	-20.3	-19.6	-20.6	-18.7
Depth (km)	Amphibole (Ridolfi, 2021)			7.40	9.70
	Biotite-only thermobarometric method (Li and Zhang, 2022)	11.9	15.1	11.1	21.2
H ₂ O (wt.%)	Amphibole (Ridolfi et al., 2010)			8.0	8.8
ΔNNO	Amphibole (Ridolfi, 2021)			-0.12	0.55

Physico-chemical conditions for the emplacement of porphyritic granite

Feldspar compositions in granitic magma can reveal the crystallizing temperature of magma (Elkins and Grove, 1990; Deer et al., 1992). Plagioclase composition on the feldspar ternary diagram (Figure 4), suggests low-temperature crystallization (<600 °C) and its sodic nature (albite to oligoclase), consistent with a highly differentiated melt (Dall'Agnol, 1982).

Biotite, the major ferromagnesian phase, can be used to understand the physico-chemical conditions of this granite during the emplacement (Henry et al., 2005; Uchida et al., 2007; Ridolfi et al., 2010). Based on the machine learning empirical formula of the biotite-only geothermobarometer (Li and Zhang, 2022), the estimated crystallization temperature-pressure for porphyritic granite yields temperature estimates in the range of 738–769 °C (average 756 °C) and 3.2–4.0 kbar (average 3.6 kbar), respectively. The calculated pressure of biotite was used to estimate the intrusion depth of the studied granitoids using the equation $P = \rho gh$, where P is crystallization pressure (GPa), ρ is the density of continental crust (2.73 g/cm³), g is the acceleration due to gravity (~9.81 m/s²), and h is depth (km). By applying this method, the calculated crystallization depth ranges from 11.7 to 14.8 km (average 13 km), corresponding to upper-crustal crystallization levels (Table 2). The calculated fO_2 for the biotite composition of porphyritic granite, based on the scheme of Wones (1989), ranges from -20.3 to -19.6 (Table 2). On the Fe²⁺-Fe³⁺-Mg ternary diagram (Figure 6d), the biotites from porphyritic granite are distributed in the NNO buffer, consistent with the magnetite series of Ishihara (1977), corresponding to oxidizing conditions.

Physico-chemical conditions for the emplacement of equigranular granite

Feldspar thermometry can be applied to estimate the temperature conditions of the crystallizing magma (Elkins and Grove, 1990; Deer et al., 1992). On the Ab-Or-An ternary diagram (Figure 4), plagioclase composition in equigranular granite indicates low-temperature crystallization (<600 °C) and a sodic range from oligoclase to andesine, consistent with a highly differentiated melt (Dall'Agnol, 1982).

The presence of titanite in a rock corresponds to the stable intercrystalline phases for Ti-saturation (Otten, 1984), and Ti content in amphibole is a function of magma crystallization temperature (Otten, 1984; Ernst and Liu, 1998; Ridolfi et al., 2010). Applying the amphibole chemistry thermobarometer (Ridolfi, 2021), the calculated temperature for equigranular granite ranges from 723 to 741 °C, with an average of 732 °C, and the pressure ranges from 1.97 to 2.61 kbar, with an average of 2.25 kbar. The Al-in-hornblende geobarometry method of Anderson and Smith (1995) is also used to validate the pressure for equigranular granite, yielding estimated values ranging from 2.09 to 2.82 kbar with an average of 2.37 kbar (Table 2). Amphibole with high Mg and low Al contents in this rock indicates low-pressure and upper crustal levels of crystallization corresponding to the emplacement depth of 7.4–9.7 km for this rock (Appendix 4). Additionally, biotite compositions of equigranular granite were used to calculate the pressure and temperature using the thermobarometric model of Li and Zhang (2022). The calculated results show temperature ranges of 769–798 °C (average 782 °C) and pressure ranges of 2.96–5.69 kbar (average 4.04 kbar) (Appendix 5). The pressure variation reflects polybaric crystallization during the magma ascent or decompression, driven by amphibole crystallization at shallower pressures than biotite crystallization (Papoutsas et al., 2016; Li and Zhang, 2022). However, it is shown that in granitic magma, amphibole (hornblende) commonly crystallizes before biotite during cooling and fractionation (Huang et al., 2019). Water content in granitic rocks

significantly influences the crystallization stability of amphibole in the melt (Naney, 1983; Dall'Agnol et al., 1999; Parat et al., 2008; de Oliveira et al., 2010). The water (H₂O) contents from amphibole compositions in equigranular granite were estimated by using the method of Ridolfi et al. (2010) with an uncertainty range of ± 1 wt.%, and the result exhibits high water contents of ~ 8 wt.% (Table 2).

The oxygen fugacity (fO_2) of magma is primarily influenced by characteristics of the magma source and the extent of magmatic differentiation (Wones and Eugster, 1965; Anderson and Smith, 1995; Ridolfi et al., 2010). The abundance of amphibole, biotite, titanite, and opaques (Fe-Ti oxides) in granitic rocks, contributes to elevating the oxygen fugacity (Czamanske and Wones, 1973; Czamanske et al., 1981). The fO_2 was calculated with amphibole composition based on the method of Ridolfi (2021), which yields $\log fO_2$ values in equigranular granite ranging from -15.6 to -14.7 (Appendix 4). The equigranular granite amphiboles are distributed in the intermediate fO_2 field and confined to values above the nickel-nickel-oxygen (NNO) buffer, respectively (Figure 5b, c). Their ΔNNO values are high, ranging between -0.12 and 0.55 (Appendix 4). Based on biotite chemistry, on the Fe^{2+} - Fe^{3+} -Mg ternary biotite diagram (Figure 6d), the equigranular granite biotite varies between NNO and hematite-magnetite (HM) buffers. They are consistent with the magnetite series of Ishihara (1977), corresponding to the oxidizing conditions (Figure 6d). The calculated crystallization depth from biotite chemistry ranges from 11 to 21 km (average 14 km) (Table 2), which corresponds to the shallow emplacement depth probably at the upper continental crustal level (Rudnick and Gao, 2003). The results from biotite and amphibole chemistry of the studied granite indicate that amphibole possibly crystallized and re-equilibrated at shallow levels, whereas biotite crystallized at relatively deeper conditions and preserved ~ 4 kbar pressure during the magma ascent. As a result, we considered that biotite pressures (~ 4 kbar) and depth of crystallization at 14 km to be more representative of the peak crystallization.

Source and tectonic setting

Amphibole and biotite are common ferromagnesian minerals in felsic rocks that can be widely used as petrogenetic indicators, and their compositions serve as a proxy indicating the nature of host magma (Zhou, 1986; Abdel-Rahman, 1994; Nachit et al., 2005; Bell et al., 2017; Samadi et al., 2021). The biotite compositions of both porphyritic granite and

equigranular granite were used to determine the source and tectonic setting (Zhou, 1986; Abdel-Rahman, 1994; Ridolfi et al., 2010; Gion et al., 2022). On the biotite tectonic discrimination diagrams of Abdel-Rahman (1994) modified by Gion et al. (2022), all the biotites from the studied granitoids are distributed in the alkaline A-type granite fields (Figure 7a). These biotite discrimination diagrams of Abdel-Rahman (1994) and Gion et al. (2022), have been treated cautiously because biotite chemistry can be modified during the re-equilibration or metamorphism. The Fe/Mg ratio and Ti contents in biotite are particularly sensitive to post-crystallization modification. However, in the present study, the biotite grains exhibit texturally well-preserved characteristics, showing limited evidence of secondary alteration, suggesting that their compositions broadly retain primary magmatic characteristics. Therefore, these diagrams for the studied granitoids are used to support the petrographic observations and tectonic constraints. On the MgO vs. FeO^T (wt.%) diagram (Figure 7b), all the biotites fall within the anorogenic alkaline suites field, except a few show proclivity towards calc-alkaline orogenic suites (Abdel-Rahman, 1994). Alkaline felsic magmatism is associated with different tectonic environments, particularly in continental rifts, syn-collisional, and post-collisional stages of continental collision (Goodenough et al., 2021).

In the $FeO^T/(FeO^T+MgO)$ vs. MgO biotite source diagram (Figure 7c), all the biotites of porphyritic granite and equigranular granite are distributed in the crustal source fields (Zhou, 1986). Amphiboles in equigranular granites also have high concentrations of TiO₂ and Al^{IV} (Appendix 4). These amphiboles have high Fe/(Fe+Mg) ratios (>0.65), suggesting an affinity for ferroan, anorogenic (A-type) granite, likely crystallized under relatively low to moderate oxygen fugacity conditions. Additionally, amphibole and biotite have $\log fO_2$ ranges of -15.6 to -14.7 and -20.6 to -18.7 , respectively, suggesting that amphibole crystallized at an intermediate stage and the biotite crystallized later. The presence of mafic enclaves and the occurrence of a hornblende, biotite, and titanite mineral assemblage in equigranular granite suggest that the crustal source interacted with mantle-derived mafic magma. Biotite compositions of equigranular granite suggest that the granite was derived from an alkaline suite in an anorogenic environment.

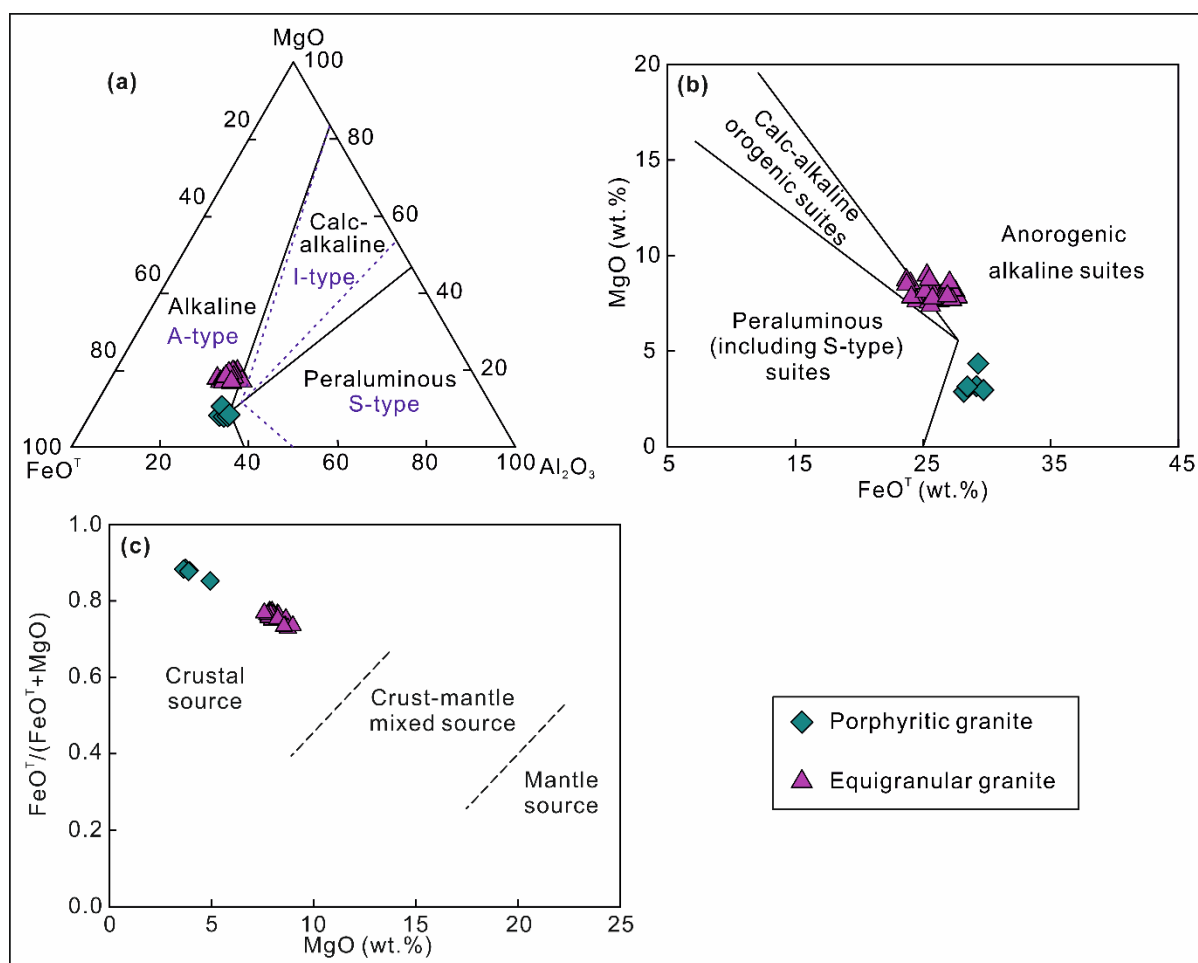


Figure 7. (a) FeO^T – MgO – Al_2O_3 ternary biotite tectonic discrimination diagram after Abdel-Rahman (1994), later modified by Gion et al. (2022), showing that the studied granitoid biotites exhibit alkaline A-type affinity. Solid black lines are after Abdel-Rahman (1994) for alkaline, calc-alkaline, and peraluminous granite fields. Dashed blue lines are new divisions after Gion et al. (2022) for A-type, I-type, and S-type granite fields; (b) MgO vs. FeO^T biotite discrimination diagram (Abdel-Rahman, 1994), displaying the anorogenic alkaline A-type suites for the studied granitoids; (c) $\text{FeO}^T/(\text{MgO} + \text{FeO}^T)$ vs. MgO biotite source diagram (Zhou, 1986), showing the crustal source for studied granitoids.

CONCLUSIONS

Based on the mineral compositions of amphibole and biotite in porphyritic granite and equigranular granite from the southwestern part of the CGGC, the following conclusions can be drawn:

- (1) Biotite thermobarometry of porphyritic granite indicates high temperatures (738–769 °C) and relatively high pressure, ranging from 3.2 to 4 kbar.
- (2) The estimated P-T conditions for equigranular granite using biotite and amphibole compositions, vary from 769 to 798 °C, 723 to 741 °C, and 3 to 5.7 kbar, 2 to 2.8 kbar, respectively.
- (3) The estimated barometric conditions for porphyritic granite suggest that it crystallized at ~13 km depth, corresponding to upper-crustal conditions.
- (4) The P-T- $f\text{O}_2$ conditions of equigranular granite indicate these granites were crystallized at a depth of ~14 km,

corresponding to upper-crustal crystallization under a high-oxidizing environment buffered in NNO (nickel-nickel-oxygen) to HM (hematite-magnetite) conditions.

(5) Biotite compositions of the studied granitoids indicate that they were formed from a crustal source due to alkaline A-type parental magma in an anorogenic tectonic environment. However, the presence of MME in equigranular granite suggests a possible limited mantle contribution to the source magma.

Acknowledgements

The authors thank the Director, CSIR-NGRI, Hyderabad for the permission to publish this work. MQ acknowledges the CSIR-JRF/ SRF fellowship. This study is carried out under MLP-7015-28 (MRM) funds. We acknowledge the EPMA facility at Banaras Hindu University, established with a SERB research grant (IR/S4/ESF-18/2011-dt. 12-11-2013). Dr. Rahul

Patel and Mr. Rakesh Kumar Patel are thanked for their help in sample collection during the fieldwork. This study forms part of MQ's doctoral thesis. The authors thank three anonymous reviewers for their constructive and valuable comments/suggestions, which substantially improved the quality of the manuscript.

Author credit statement

Mohammad Qasim: Writing the original draft, reviewing, validation, and conceptualization. Satya Narayana Mahapatro: Writing the original draft, editing, visualization, and conceptualization. Aparajita Tripathi: Validation, formal analysis, and data curation. M. Ram Mohan: Writing the original draft, editing, supervision, conceptualization, and funding acquisition.

Data availability

Data is provided in the manuscript and can be available from the corresponding author on reasonable request.

Compliance with Ethical Standards

The authors declare no conflict of interest and adhere to copyright norms.

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Received on: 11-02-2026; Revised on: 06-06-2026; Accepted on: 25-6-2026

Appendix 1. Mineral assemblages (vol.%) of granitoid samples from the southwestern part of CGGC

Rock Type	<i>Porphyritic Granite</i>	<i>Equigranular Granite</i>	
Sample No.	CG-1	CG-2	CG-4
<i>Content of mineral (vol %)</i>			
Plagioclase	18-20	16-18	18-22
Orthoclase	12-14	12-14	12-18
Microcline	20-25	10-12	18-20
Quartz	25-28	20-25	22-25
Biotite	10-12	16-18	12-15
Amphibole	-	8-10	-
Titanite	<1	1-2	1-2
Zircon	<1	1	1
Apatite	<1	<1	<1

Appendix 2. Mineral composition of K-feldspars for granitoids from the southwestern part of CGGC

Rock type	<i>Porphyritic Granite</i>												<i>Equigranular Granite</i>						
Sample No.	CG-1												CG-4						
Analysis point	46/1	50/1	51/1	52/1	53/1	54/1	56/1	61/1	62/1	70/1	71/1	81/1	103/1	105/1	106/1	107/1	118/1	119/1	120/1
<i>Major oxides (wt.%)</i>																			
SiO ₂	64.2	64.6	64.2	64.6	64.7	64.2	64.9	64.3	64.7	64.4	64.5	64.5	64.7	64.8	64.7	64.7	64.5	64.2	64.5
TiO ₂	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al ₂ O ₃	18.5	18.3	18.3	18.5	18.6	18.6	18.6	18.4	18.7	18.4	18.2	18.6	18.3	18.6	18.5	18.5	18.3	18.4	18.7
FeO ^T	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.22	0.00	0.07	0.00	0.00	0.00
MnO	0.00	0.00	0.00	0.00	0.00	0.03	0.04	0.00	0.00	0.00	0.00	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.01	0.03	0.00	0.00	0.00	0.04	0.00	0.02	0.00	0.00	0.03	0.00	0.01	0.00	0.01	0.02	0.01	0.00	0.01
Na ₂ O	0.74	0.80	0.76	0.74	0.73	0.62	0.75	0.96	1.09	0.75	0.63	0.73	0.70	0.66	0.64	0.53	0.53	0.61	0.57
K ₂ O	15.3	14.8	15.6	15.1	15.1	15.5	15.4	15.0	14.8	15.2	15.2	15.3	15.2	15.0	15.2	15.0	15.3	15.3	15.1
Total	98.7	98.6	98.9	98.9	99.1	99.0	99.6	98.7	99.4	98.7	98.5	99.2	98.9	99.2	99.1	98.8	98.7	98.5	98.8
<i>Atom numbers calculated based on 8 (O) atoms per formula unit</i>																			
Si	3.00	3.03	2.99	3.02	3.02	3.00	3.01	3.00	3.00	3.01	3.03	3.00	3.02	3.02	3.02	3.03	3.02	3.01	3.02
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	1.02	1.01	1.01	1.02	1.02	1.02	1.01	1.01	1.02	1.01	1.01	1.02	1.01	1.02	1.02	1.02	1.01	1.02	1.03
Fe ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.07	0.07	0.07	0.07	0.07	0.06	0.07	0.09	0.10	0.07	0.06	0.07	0.06	0.06	0.06	0.05	0.05	0.06	0.05
K	0.91	0.88	0.93	0.90	0.90	0.92	0.91	0.89	0.88	0.91	0.91	0.91	0.91	0.89	0.90	0.90	0.92	0.92	0.90
Total (cations)	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
<i>Nominal end-member fractions (wt.%)</i>																			
Ab	6.81	7.58	6.90	6.95	6.85	5.69	6.94	8.83	10.07	7.01	5.91	6.81	6.48	6.23	6.02	5.06	5.01	5.72	5.46
An	0.04	0.13	0.00	0.00	0.00	0.20	0.00	0.09	0.00	0.00	0.17	0.00	0.07	0.00	0.05	0.10	0.07	0.00	0.07
Or	93.2	92.3	93.1	93.0	93.2	94.1	93.1	91.1	89.9	93.0	93.9	93.2	93.5	93.8	93.9	94.8	94.9	94.3	94.5

Appendix 3. Mineral composition of plagioclase for granitoids from the southwestern part of CGGC

Rock type	Porphyritic Granite						Equigranular Granite																																		
Sample No.	CG-1						CG-2																				CG-4														
Analysis point	49/1	65/1	66/1	67/1	68/1	69/1	10/1	11/1	12/1	13/1	14/1	15/1	20/1	21/1	25/1	26/1	27/1	28/1	32/1	33/1	48/1	56/1	61/1	62/1	67/1	72/1	78/1	79/1	80/1	91/1	92/1	94/1	95/1	96/1	113/1	114/1	115/1				
Major oxides (wt.%)																																									
SiO ₂	64.6	64.8	66.1	66.9	64.9	66.0	58.1	58.4	58.0	58.4	59.0	57.4	58.5	59.8	58.3	59.0	59.1	58.5	58.6	57.3	59.0	58.0	59.5	58.2	57.5	58.2	60.3	60.6	60.2	61.9	61.3	60.2	60.3	60.3	64.1	64.0	64.0				
TiO ₂	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.00	0.00					
Al ₂ O ₃	22.0	21.5	21.1	20.1	21.9	21.6	26.8	26.3	26.7	26.4	26.3	27.7	26.7	25.8	26.8	25.8	25.8	26.3	26.4	27.1	26.5	26.1	25.8	27.1	27.4	25.8	25.0	25.0	24.9	23.8	24.4	25.2	25.1	25.1	22.9	22.3	23.1				
FeO ^T	0.00	0.00	0.00	0.00	0.00	0.00	0.08	0.13	0.07	0.00	0.03	0.08	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00				
MnO	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.12	0.01	0.03	0.00	0.00	0.00	0.00	0.03	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
CaO	2.83	2.70	1.89	0.87	2.93	2.55	8.11	7.75	8.05	8.04	7.21	8.69	8.07	7.06	8.60	7.18	7.37	7.67	7.88	8.88	7.58	7.47	7.24	8.19	8.95	7.48	6.38	6.38	6.30	5.19	5.77	6.13	6.66	6.31	3.63	3.64	3.89				
Na ₂ O	9.84	10.1	10.5	10.8	9.98	10.1	6.36	6.47	6.31	6.40	7.08	6.13	6.45	7.17	6.53	7.06	7.12	6.73	6.92	6.25	6.93	6.81	6.93	6.57	6.20	6.92	7.53	7.53	7.45	8.03	8.04	7.85	7.74	7.27	9.47	9.34	9.14				
K ₂ O	0.11	0.11	0.03	0.00	0.02	0.02	0.07	0.09	0.08	0.12	0.09	0.13	0.08	0.14	0.05	0.02	0.09	0.03	0.03	0.05	0.06	0.05	0.13	0.06	0.11	0.08	0.10	0.07	0.07	0.07	0.08	0.08	0.09	0.15	0.05	0.00	0.12				
Total	99.4	99.3	99.6	98.7	99.7	100.3	99.6	99.3	99.3	99.4	99.7	100.1	99.8	100.0	100.4	99.1	99.5	99.2	99.9	99.6	100.0	98.5	99.6	100.2	100.2	98.5	99.3	99.6	98.9	98.9	99.6	99.5	99.8	99.2	100.1	99.3	100.3				
Atom numbers calculated based on 8 (O) atoms per formula unit																																									
Si	2.86	2.87	2.91	2.98	2.87	2.90	2.62	2.64	2.62	2.64	2.64	2.57	2.63	2.67	2.61	2.66	2.66	2.64	2.62	2.58	2.64	2.64	2.67	2.60	2.58	2.64	2.71	2.71	2.72	2.78	2.74	2.69	2.69	2.72	2.82	2.85	2.82				
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00					
Al	1.15	1.12	1.09	1.05	1.14	1.12	1.42	1.40	1.42	1.41	1.39	1.46	1.41	1.36	1.41	1.37	1.36	1.40	1.40	1.44	1.40	1.40	1.37	1.43	1.45	1.38	1.32	1.32	1.32	1.26	1.28	1.33	1.32	1.33	1.19	1.17	1.20				
Fe ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00					
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00					
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00					
Ca	0.13	0.13	0.09	0.04	0.14	0.12	0.39	0.38	0.39	0.39	0.35	0.42	0.39	0.34	0.41	0.35	0.35	0.37	0.38	0.43	0.36	0.36	0.35	0.39	0.43	0.36	0.31	0.31	0.30	0.25	0.28	0.29	0.32	0.30	0.17	0.17	0.18				
Na	0.85	0.87	0.90	0.93	0.85	0.86	0.56	0.57	0.55	0.56	0.62	0.53	0.56	0.62	0.57	0.62	0.62	0.59	0.60	0.55	0.60	0.60	0.60	0.57	0.54	0.61	0.66	0.65	0.65	0.70	0.70	0.68	0.67	0.63	0.81	0.81	0.78				
K	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01				
Total (cations)	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00				
Nominal end-member fractions (wt.%)																																									
Ab	85.7	86.6	90.8	95.7	86.0	87.7	58.4	59.8	58.4	58.6	63.7	55.7	58.8	64.2	57.7	64.0	63.3	61.2	61.3	55.9	62.1	62.1	62.9	59.0	55.3	62.3	67.7	67.8	67.9	73.4	71.2	69.5	67.4	67.0	82.3	82.3	80.4				
An	13.6	12.8	9.03	4.28	13.9	12.2	41.2	39.6	41.1	40.7	35.8	43.6	40.7	35.0	42.0	35.9	36.2	38.6	38.6	43.9	37.5	37.6	36.3	40.7	44.1	37.2	31.7	31.8	31.7	26.2	28.3	30.0	32.1	32.1	17.4	17.7	18.9				
Or	0.61	0.62	0.16	0.00	0.12	0.12	0.42	0.55	0.48	0.74	0.53	0.77	0.50	0.81	0.28	0.11	0.51	0.21	0.18	0.27	0.36	0.28	0.78	0.36	0.63	0.46	0.59	0.40	0.39	0.41	0.49	0.46	0.50	0.94	0.30	0.00	0.70				

Appendix 4. Mineral chemistry of amphibole for granitoids from the southwestern part of CGGC

Rock types	<i>Equigranular Granite</i>						
Sample	CG-2						
Analysis point	22/1	23/1	36/1	37/1	38/1	39/1	40/1
<i>Major oxides (wt.%)</i>							
SiO ₂	42.3	42.8	42.0	42.0	42.2	43.0	42.3
P ₂ O ₅	0.11	0.06	0.10	0.06	0.04	0.08	0.08
TiO ₂	1.00	1.02	0.65	0.81	1.16	1.07	1.05
Al ₂ O ₃	10.5	10.6	11.5	10.6	10.9	10.5	11.1
Cr ₂ O ₃	0.04	0.00	0.05	0.07	0.04	0.02	0.05
MnO	0.52	0.46	0.45	0.51	0.30	0.37	0.44
FeO ^T	21.9	21.3	21.5	21.4	20.5	21.4	21.5
NiO	0.04	0.16	0.06	0.10	0.14	0.10	0.06
MgO	6.24	6.16	6.12	6.36	6.38	6.75	6.32
CaO	11.4	10.9	11.3	11.1	11.4	11.6	11.3
Na ₂ O	1.01	1.04	1.00	1.05	1.01	0.98	0.98
K ₂ O	1.32	1.41	1.35	1.32	1.32	1.23	1.40
H ₂ O ⁺	1.78	1.72	1.75	1.68	1.72	1.70	1.57
F	0.27	0.41	0.32	0.45	0.39	0.45	0.70
Cl	0.16	0.17	0.21	0.20	0.19	0.18	0.20
Total	98.6	98.2	98.4	98.0	97.7	99.4	99.1
<i>Atom numbers calculated based on 23 (O) atoms per formula unit</i>							
<i>T (ideally 8 apfu)</i>							
Si	6.61	6.69	6.56	6.61	6.62	6.63	6.58
P	0.007	0.004	0.006	0.004	0.003	0.005	0.005
Be	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	1.39	1.31	1.43	1.39	1.38	1.36	1.42
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00
T subtotal	8.00	8.00	8.00	8.00	8.00	8.00	8.00
<i>C (ideally 5 apfu)</i>							
Ti	0.12	0.12	0.08	0.10	0.14	0.12	0.12
Al	0.55	0.64	0.68	0.58	0.63	0.55	0.61
Cr	0.005	0.000	0.006	0.008	0.006	0.003	0.006
Fe ³⁺	0.11	0.09	0.12	0.12	0.02	0.08	0.11
Ni	0.005	0.020	0.008	0.013	0.018	0.012	0.008
Mn ²⁺	0.009	0.012	0.000	0.000	0.027	0.000	0.000
Fe ²⁺	2.76	2.69	2.69	2.69	2.67	2.67	2.69

Mg	1.45	1.43	1.43	1.49	1.49	1.55	1.47
C subtotal	5.00	5.00	5.00	5.00	5.00	5.00	5.00
<i>B (ideally 2 apfu)</i>							
Mn ²⁺	0.06	0.05	0.06	0.07	0.01	0.05	0.06
Fe ²⁺	0.000	0.000	0.006	0.003	0.000	0.000	0.004
Ca	1.90	1.82	1.89	1.88	1.92	1.92	1.88
Na	0.04	0.13	0.05	0.05	0.06	0.03	0.06
B subtotal	2.00	2.00	2.00	2.00	2.00	2.00	2.00
<i>A (from 0 to 1 apfu)</i>							
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Li	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.27	0.19	0.25	0.27	0.24	0.26	0.24
Pb	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K	0.26	0.28	0.27	0.26	0.27	0.24	0.28
A subtotal	0.53	0.47	0.52	0.53	0.51	0.50	0.51
O (non-W)	22.0	22.0	22.0	22.0	22.0	22.0	22.0
<i>W (ideally 2 apfu)</i>							
OH	1.82	1.75	1.79	1.72	1.76	1.73	1.60
F	0.13	0.20	0.16	0.22	0.19	0.22	0.34
Cl	0.04	0.05	0.05	0.05	0.05	0.05	0.05
W subtotal	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Sum of T, C, B, A	15.5	15.5	15.5	15.5	15.5	15.5	15.5
Total Al	1.93	1.95	2.11	1.97	2.02	1.91	2.03
Total Fe	2.86	2.78	2.81	2.82	2.69	2.76	2.80
Mg/(Mg+Fe)	0.51	0.52	0.51	0.53	0.55	0.56	0.52
Fe/(Fe+Mg)	0.66	0.66	0.66	0.65	0.64	0.64	0.66
Mg/(Mg+Fe ²⁺)	0.34	0.35	0.35	0.36	0.36	0.37	0.35
Ca+Na+K	2.47	2.42	2.46	2.46	2.50	2.46	2.45
T (°C) (Ridolfi, 2021)	737	725	734	741	724	723	740
P (kbar) (Ridolfi, 2021)	2.10	2.19	2.61	2.25	2.26	1.97	2.39
Log fO ₂	-15.2	-15.6	-14.7	-14.8	-15.5	-15.5	-14.8
Depth (km) (Ridolfi, 2021)	7.9	8.2	9.7	8.4	8.5	7.4	8.9
H ₂ O (wt.%) (Ridolfi et al., 2010)	8.0	8.1	8.8	8.1	8.6	8.2	8.2
ΔNNO	0.04	-0.12	0.55	0.34	0.05	0.08	0.30

Behaviour of surface temperature during high-magnitude earthquakes ($M \geq 6.0$)

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ABSTRACT

As the Earth prepares itself before an earthquake, different anomalous signs in the atmospheric parameters, such as surface temperature, are generally observed. A thorough examination of these anomalous signs assists us in understanding the earthquake mechanisms and seismic predictions. In the present work, the surface thermal parameter - surface temperature (ST) during the twelve recent earthquakes of high magnitude ($M \geq 6.0$) are studied. All the earthquakes considered in the study originated on land. The data of surface temperature have been retrieved from the NCEP/NCAR website re-analysis. The value of ST considered is one month pre- and post- to the earthquakes. It is observed that all the earthquakes show an anomaly during the seismic-affected periods (pre- and post). Most of the earthquakes considered show a maximum anomalous surge in ST within $\sim 1-9$ days prior to the mainshock. Also, post-seismic anomalies are observed during most of the events. In some of the events, the maximum anomalies occurred on the day of the earthquake. The anomalous behaviour of normalised ST varied from $\sim 6.15\%$ to 116.6% as compared to background noise during the pre-seismic period of the considered earthquakes. During the post-seismic period, this variation spans from $\sim 17.69\%$ to 83.95% compared to the background noise.

Keywords: Surface temperature (ST) anomaly, Earthquake, Magnitude, Precursors, Atmospheric parameters, Earthquake preparation zone (EPZ).

INTRODUCTION

Earthquakes are among the most destructive natural disasters, striking without warning and highlighting the urgent need for effective prediction methods to mitigate their impact. Despite of ongoing research, accurate forecasting of seismic events, remains a significant challenge due to their complex nature. Prior to a strong-magnitude earthquake, the Earth undergoes preparatory processes during which multi-parametric precursors are recorded within a timeframe and specific region known as Earthquake Preparation Zone (EPZ), introduced by Dobrovolsky et al. (1979). Various precursory studies that encompass ionospheric precursors (Pulinets and Boyarchuk, 2004), mechanical stress changes (Niu, 2008), electromagnetic signals (Kondo, 1966), hydrological precursors (Rikitake, 1979) and thermal irregularities (Pulinets et al., 2006; Bardhan et al., 2022; Sharma et al., 2024) have been carried out. Thermal anomalies within earthquake preparation zones, have been observed several days before major shocks (Tronin et al., 2002) and are often associated with increased thermal flux originating from the Earth's crust in tectonically active regions (Ouzounov and Freund, 2004). These anomalies, frequently detected through satellite-based thermal infrared (TIR) sensors, manifest as enhanced TIR emissions from the Earth's surface prior to seismic events and are collectively referred to as thermal anomalies (Freund et al., 2005). Surface Temperature (ST), a key parameter, reflects the thermal state of the Earth's surface and is commonly monitored using satellite TIR observations.

Thermal precursor detection methodologies primarily involve comparing land surface temperatures (LST) from the days leading up to an earthquake with temperatures from the same season in a non-seismic year (Bhardwaj et al., 2017). Different

studies have identified LST anomalies appearing at various time intervals before an earthquake, ranging widely from 2 to 24 days. Ouzounov et al. (2006) reported LST anomalies occurring 4 to 14 days prior to seismic events and Tronin et al. (2002) observed anomalies within a window of 6 to 4 days before the earthquake. Similarly, Boudriki Senglali et al. (2024) analyzed thousands of earthquake events with magnitudes greater than 4 ($M > 4$) using GOES and AQUA satellite observations and found that the LST increased before the earthquakes and thereafter decreased. Studies have shown that LST can exhibit anomalous increases, typically between 1°C and 5°C , several days to weeks before major earthquakes (Tronin et al., 2002; Pulinets and Ouzounov, 2011). Tibet's 2025 Dingri study confirmed MODIS LST anomalies (STL/RST validated) up to 6 months pre-event, fault-aligned (Li et al., 2025). These anomalies are believed to be triggered by underlying processes such as stress-induced deformation of rocks, the release of geothermal energy, and the emission of gases like radon, which collectively enhance the thermal conductivity of surface materials. Presently the study focuses on thermal precursors-temperature anomalies near the surface that appear prior to seismic events. This study examines the potential of Surface Temperature (ST) as a thermal precursor by analysing its behaviour before and after twelve high-magnitude earthquakes ($M \geq 6.0$) across various global regions.

DATA SELECTION AND SURFACE TEMPERATURE ANALYSIS

In this study, the twelve high-magnitude earthquakes were analyzed ($M \geq 6.0$). Comprehensive information such as event location, date, time, epicentre coordinates and focal depth was obtained from the National Earthquake Information

Center (NEIC) and the United States Geological Survey (USGS) website: <http://earthquake.usgs.gov/activity/world.html>. The details of selected earthquake events are listed in Table 1. ST data corresponding to twelve recent high-magnitude earthquakes were obtained from the National Centers for Environmental Prediction/National Center for

Atmospheric Research (NCEP-NCAR). The dataset covers the Earthquake Preparation Zones (EPZ), including the epicentres of the selected seismic events. The daily ST data were retrieved from the link: <https://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP-NCAR/.CDAS-1/.DAILY/.Diagnostic/.surface/.temp/datasetselection.html?Set-Language=en>.

Table 1. Details of the studied global earthquake events

S. no.	Location	Date & Time (UTC)	Epicentre (Lat., Long.)	Magnitude	Focal depth (km)	Background noise $ST(\mu) + 1.5\sigma$
1	Southern Xinjiang, China	22.01.2024 (18:09:04)	41.25°N, 78.65°E	7.0	13	1.24 (Jan), 1.62 (Feb)
2	Western Afghanistan	07.10.2023 (06:41:03)	34.59°N, 61.92°E	6.3	14	2.50 (Sep), 0.48 (Oct)
3	Eastern Turkey	06.02.2023 (10:26:46)	38.03°N, 38.09°E	6.0	10	0.71 (Jan), 1.87 (Feb)
4	Western Sichuan China	05.09.2022 (04:52:19)	29.67°N, 102.23°E	6.6	12	1.88 (Aug), 1.50 (Sep)
5	Yunnan, China	21.05.2021 (13:48:37)	25.72°N, 100.00°E	6.1	9	2.06 (May), 3.45 (Jun)
6	Assam, India	28.04.2021 (02:21:26)	26.78°N, 92.45°E	6.0	34	1.03 (Apr), 1.53 (May)
7	Hotan, China	25.06.2020 (21:05:18)	35.59°N, 82.41°E	6.3	10	0.46 (Jun), 0.73 (Jul)
8	Mindanao, Philippines	31.10.2019 (01:11:19)	6.91°N, 125.17°E	6.5	10	1.06 (Oct), 0.59 (Nov)
9	Mindanao, Philippines	29.10.2019 (01:04:43)	6.75°N, 125.00°E	6.6	15	1.09 (Oct), 0.67 (Nov)
10	Breñón, Panama	12.05.2019 (19:24:50)	8.62°N, 82.83°W	6.0	19	0.71 (Apr), 1.31 (May)
11	Luzon, Philippines	22.04.2019 (09:11:12)	14.95°N, 120.51°E	6.1	21.8	1.36 (Apr), 1.67 (May)
12	Shi Yomi, India	17.11.2017 (22:34:19)	29.83°N, 94.98°E	6.4	8	2.74 (Oct), 1.44 (Nov)

The daily surface temperature (ST_d) values were considered over a 60-day window, encompassing both the pre- and post-earthquake periods with the earthquake day as the reference point. Thereafter, mean value (ST_{mean}) was computed over the window of 60 days.

$$ST_{mean} = \frac{1}{60} \sum_{d=1}^{60} ST_d \quad \text{-----} \quad (1)$$

Where $d=1, 2, \dots, 60$ represent each day within the pre- and post- earthquake observation window, ST_d denotes the surface temperature recorded on day d . To remove seasonal and climatological effects, the surface temperature anomaly (ΔST) was calculated by subtracting the long-term monthly mean surface temperature for each day from the corresponding daily values:

$$\Delta ST = ST_d - ST_{monthly\ mean} \quad \text{-----} \quad (2)$$

Here $ST_{monthly\ mean}$ represents the monthly mean surface temperature for the corresponding day d , calculated from 10-year historical dataset.

For normalization, the surface temperature anomaly (ΔST) was divided by the standard deviation (σ) of the corresponding day, calculated from a 10-year surface temperature dataset.

$$\Delta ST_{normalised} = \frac{\Delta ST}{\sigma} \quad \text{-----} \quad (3)$$

This normalization allowed consistent comparison of surface temperature variations across different time periods and geographic locations.

To identify the variation in ST as potential anomalies, background noise has been calculated. The background noise is calculated as the monthly mean of normalized surface temperature obtained from equation (3) and is represented as $ST(\mu)$ plus 1.5 times the standard deviation (σ).

$$\text{Background Noise} = ST(\mu) + 1.5\sigma \quad \text{-----} \quad (4)$$

Any normalized ΔST value that exceeded this background noise is classified as anomalous and potentially reflective of earthquake activity. Also, the present data analysis accounts for meteorological effects. Equation 1 represents the 60-day mean of a thermal parameter surface temperature, where ST_d is the daily value on day d . This averaging reduces the influence of short-term day-to-day weather fluctuations and provides a steady background thermal state over the analysis window encompassing both pre- and post- earthquake periods. In eq. 2, the daily value is subtracted with the long-term monthly mean derived from a 10-year dataset. This step removes seasonal and climatological variability, such as monsoon effects, and regional climate patterns, which are the dominant meteorological controls on surface temperature. Using eq. 3, the anomaly is normalized using the standard deviation calculated from a 10-year daily dataset. This accounts for natural meteorological variability by scaling the anomaly relative to typical weather-driven fluctuations.

The equation 4 computes the background noise and anomalies exceeding this level, are statistically significant and unlikely to be caused solely by predictable meteorological processes. Together, these steps explicitly account for meteorological

effects by removing seasonal trends, minimizing daily weather noise, and applying a statistically robust threshold based on long-term variability. As a result, the remaining anomalies represent departures from normal atmospheric behavior and are therefore more likely associated with earthquake-related processes rather than meteorological influences.

RESULTS AND DISCUSSION

In this study, twelve recent high-magnitude earthquakes ($M \geq 6.0$) that occurred between 2017 to 2024 have been investigated, focusing on the behavior of ST as a potential earthquake precursor, as well as post-seismic thermal response. Figure 1a showed anomalous ST variation during earthquake that occurred on January 22, 2024, in Southern Xinjiang, China. Figure 1a illustrates the variation of normalized surface temperature (ST) anomalies over the span of 60-day including the earthquake day. The figure reveals notable anomalous behaviour in surface temperature behaviour before and after the seismic event. In the figure, the black line represents the normalised daily ST values, while the red line with star indicates the background noise, defined as the mean plus 1.5 times the standard deviation ($\mu + 1.5\sigma$). The maximum background noise corresponding to each earthquake event is presented in Table 1. A pre-seismic anomaly is observed 8th

day prior to the event, with normalized ST exceeding background noise by 20.96%. In contrast, a stronger post-seismic anomaly occurred 24 days after the earthquake, showing an 83.95% increase, indicating significantly enhanced thermal activity in the post-seismic phase. Figure 1b (earthquake on October 7, 2023, Western Afghanistan) depicts the normalized surface temperature (ST), showing distinct anomalous behaviour prior to the event. The precursory analysis reveals that the normalized ST values exceed the background noise threshold on several distinct days, which are considered as anomalous. A pre-seismic anomaly is observed 2 days before the event, where normalized ST exceeding background noise by 116.67%, indicating a clear thermal signal during the earthquake preparation phase. The third earthquake analyzed in this study occurred on February 6, 2023, in Eastern Turkey (Figure 1c). Pre-seismic anomaly was observed 9 days before the event with a 39.47% increase, while a stronger post-seismic anomaly 21 days after the earthquake showed a 50.80% increase, indicating greater thermal activity in the post-seismic phase. The fourth earthquake occurred (earthquake during September 5, 2022, Western Sichuan, China) (Figure 1d) shows a post-seismic anomaly 10 days after the event, with normalized ST exceeding background noise by 46.66%, indicating a notable thermal increase.

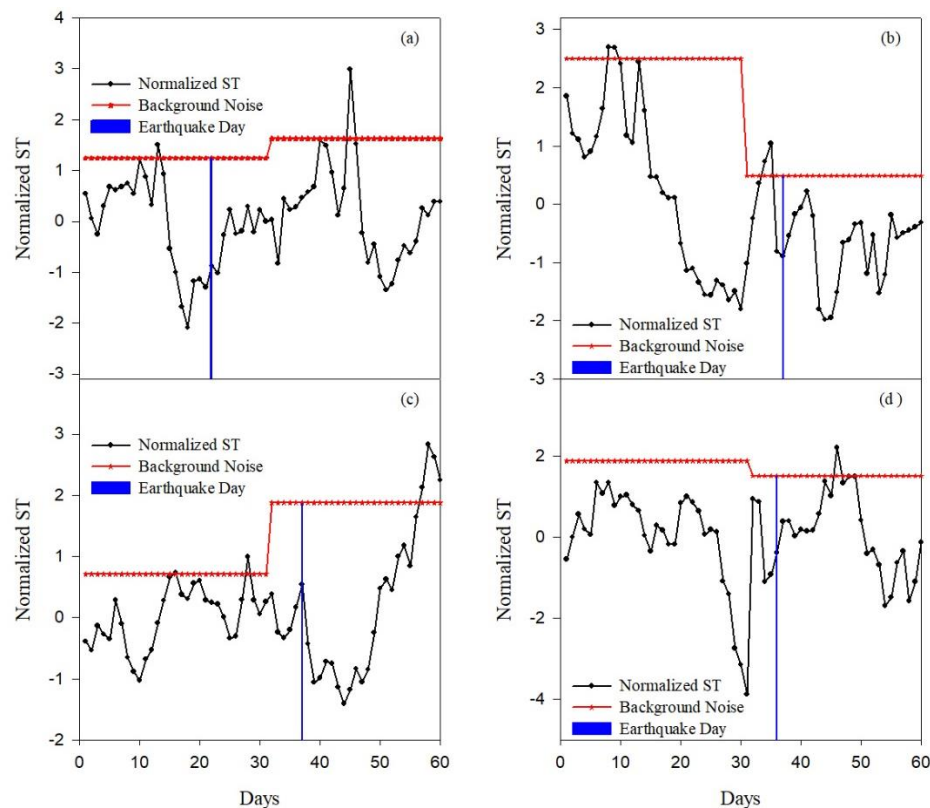


Figure 1. Representation of normalized ST during the, (a) Southern Xinjiang, China (22.01.2024), (b) Western Afghanistan (07.10.2023), (c) Eastern Turkey (06.02.2023) and (d) Western Sichuan, China (05.09.2022) earthquakes. The red horizontal line with stars represents the background noise and the vertical blue line shows the earthquake day.

The earthquake analysed that occurred on May 21, 2021, in Yunnan, China is depicted in Figure 2a. Analysis shows normalized ST exceeded background noise on multiple days, with a strong pre-seismic anomaly on the earthquake day showing a 72.81% increase, while a post-seismic anomaly one day after showed a 23.30% increase, indicating stronger thermal activity in the pre-seismic phase. Figure 2b (earthquake on April 28, 2021, Assam, India) shows the pre-seismic anomaly 1 day before the earthquake showed a 27.18% increase, while a post-seismic anomaly 23 days after showed a 26.14% increase, indicating comparable thermal variations in both phases. The next earthquake, occurred on June 25, 2020, in Hotan, China as represented by Figure 2c. It shows the pre-seismic anomaly 1 day before the earthquake showed a 76.08% increase, while a post-seismic anomaly 32 days after showed a 39.72% increase, indicating a stronger pre-seismic thermal signal. In The next earthquake event considered on October 31, 2019, in Mindanao, Philippines (Figure 2d). The precursory analysis reveals that the normalized ST values exceed the background noise threshold on several days, marking them as anomalous. A pre-seismic anomaly 1 day before the event showed a 16.03% increase, indicating a thermal deviation above background noise.

Similarly, the next earthquake analyzed in this study occurred on October 29, 2019, in Mindanao, Philippines (Figure 3a). The precursory analysis of Figure 3a shows that the normalized ST values exceed the background noise threshold on several days, indicating the presence of anomalies.

A pre-seismic anomaly on the earthquake day showed a 15.59% increase, while a post-seismic anomaly 1 day after showed a 24.77% increase, indicating a stronger post-seismic thermal response. In Figure 3b (earthquake on May 12, 2019, Breñón, Panama) shows pre-seismic anomaly on the earthquake day, showed a 6.15% increase, while a post-seismic anomaly 1 day after showed a 17.69% increase, indicating a stronger post-seismic thermal response. Figure 3c (earthquake event on April 22, 2019, in Luzon, Philippines) shows that 6 days after the event exhibited a 42.64% increase, indicating a notable thermal deviation above background noise. In a similar manner, Figure 3d (earthquake event on November 17, 2017, Shi Yomi, India) shows a pre-seismic anomaly, detected one day before the event, where the normalized ST exceeded background noise by 38.88%, indicating a significant thermal anomaly.

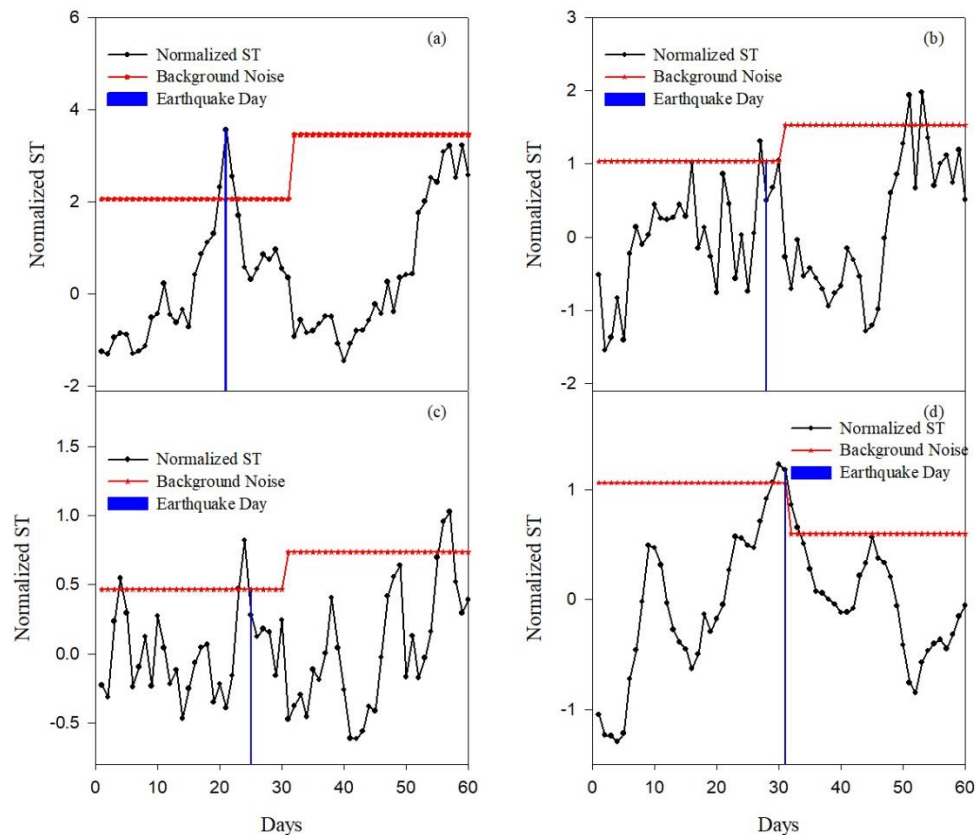


Figure 2. Representation of normalized ST during the, (a) Yunnan, China (21.05.2021), (b) Assam, India (28.04.2021), (c) Hotan, China (25.06.2020) and (d) Mindanao, Philippines (31.10.2019) earthquakes. The red horizontal line with stars represents background noise and the vertical blue line shows the earthquake day.

All the earthquakes showed significant surface thermal anomalies, during pre-seismic periods. This increase in ΔST ranged from $\sim 6.15\%$ to 116.67% above background noise and occurred between 1-9 days before the mainshock. Post-seismic anomalies are also evident in several events, with percentage increase ranging from $\sim 17.69\%$ to 83.95% . The normalized ΔST values generally remained within the defined background noise range throughout most of the observation period, in each event, they surpassed this range just before and after few days of the earthquake day. The anomalous behavior of normalized ST during the pre- and post-earthquake periods are summarised in Table 2.

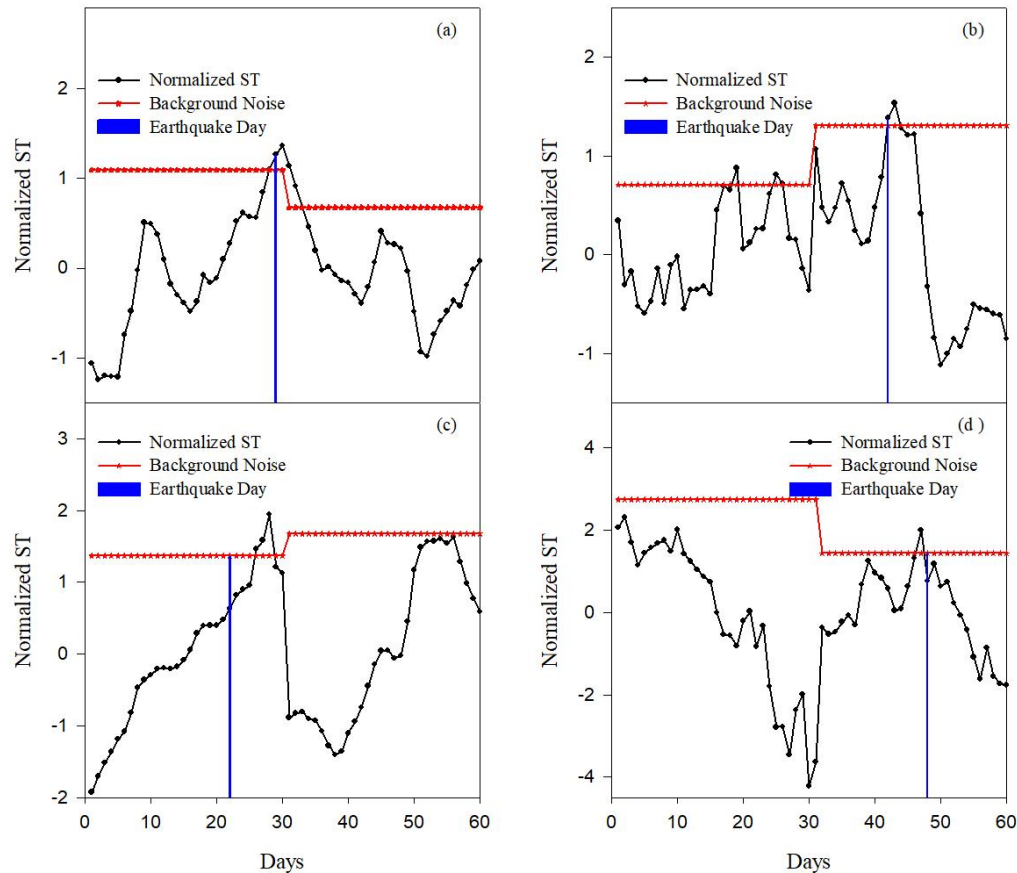


Figure 3. Representation of normalized ST during the, (a) Mindanao, Philippines (29.10.2019), (b) Breñón, Panama (12.05.2019), (c) Luzon, Philippines (22.04.2019) and (d) Shi Yomi, India (17.11.2017) earthquakes. The red horizontal line with stars represents background noise and the vertical blue line shows the earthquake day.

Table 2. Maximum normalized ST anomalies observed during the pre and post- seismic periods for all selected earthquake events.

S. no.	Location	Pre-seismic anomaly (Days before EQ)	Pre-seismic increases (%)	Post-seismic anomaly (Days after EQ)	Post-seismic increases (%)
1	Southern Xinjiang, China	8 days	20.96	24 days	83.95
2	Western Afghanistan	2 days	116.6	No anomaly	No anomaly
3	Eastern Turkey	9 days	39.47	21 days	50.80
4	Western Sichuan China	No anomaly	No anomaly	10 days	46.66
5	Yunnan, China	EQ day	72.81	1 day	23.30
6	Assam, India	1 day	27.18	23 days	26.18
7	Hotan, China	1 day	76.08	32 days	39.72
8	Mindanao, Philippines	1 day	16.03	No anomaly	No anomaly
9	Mindanao, Philippines	EQ day	15.59	1 day	24.77
10	Breñón, Panama	EQ day	6.15	1 day	17.69
11	Luzon, Philippines	No anomaly	No anomaly	6 day	42.64
12	Shi Yomi, India	1 day	38.88	No anomaly	No anomaly

This consistent exceedance across all the events highlights the potential of ST as a reliable thermal precursor during major earthquakes. It is well established that ST is influenced by a range of climatological factors, including relative humidity, atmospheric moisture content, ambient temperature, measurement location, seasonal variations, monsoon activity, local wind patterns and tidal influences. The seismic events analysed in this study originate from diverse geographic regions on land. While anomalous ST signatures were observed in each event, the nature and timing of these anomalies varied across different locations. This variability is likely attributed to differences in regional geographic and meteorological conditions. For instance, there is significant decrease in surface temperature prior to the two earthquake events (Figures 1a and 2a) which may be attributed to changes in near-surface energy balance associated with earthquake preparation processes. During the stress accumulation phase, enhanced micro-fracturing in the crust can lead to the release of gases and increased soil moisture, which promotes evaporative cooling at the surface. This process can result in a temporary reduction in surface temperature immediately before the seismic event. The proximity to large water bodies can significantly affect local relative humidity and water vapor content.

Identifying potential thermal anomalies within ST datasets necessitates establishing a reliable baseline to evaluate deviations. Several mechanisms are thought to contribute to the rise in land surface temperature observed prior to earthquakes. The earthquake phenomenon originates from depths ranging from a few to several kilometers beneath the Earth's surface. Continuous tectonic plate movement leads to the gradual buildup of stress within rock formations, ultimately causing them to rupture. This sudden rupture releases a large amount of energy, generating seismic waves that travel toward the Earth's surface (Scholz, 2019). Prior to the occurrence of earthquakes, the build up of stress within rocks is believed to be a primary factor contributing to the rise in land surface temperature. This accumulated stress is thought to generate thermal infrared emissions, which in turn enhance thermal activity near the surface (Ouzounov and Freund, 2004). Another proposed mechanism involves the formation of micro-pores in the Earth's crust due to increasing tectonic pressure. Radon, a naturally occurring radioactive gas present in soil and air, is also considered a significant precursor to earthquakes. Studies suggest that radon undergoes ionization in the period leading up to seismic events, which contributes to increased atmospheric humidity and, subsequently, a rise in surface temperature (Pulinets et al., 2006).

In addition to ongoing studies on the existence of thermal precursors, their timing and quantification have also varied significantly across studies. For instance, Ouzounov et al. (2006) observed LST anomalies occurring 4 to 14 days before

seismic events, whereas Tronin et al. (2002) reported anomalies appearing between 6 and 4 days prior to an earthquake. This variability highlights the need for further investigation into the underlying causes and mechanisms of such anomalies. All the events we analyzed presented pronounced ST anomalies, denoting remarkable deviations from baseline conditions. These anomalies were observed during both the pre-seismic and post-seismic periods surrounding the day of the earthquake.

Further, the Lithosphere-Atmosphere-Ionosphere Coupling is probably one of the most complex and interesting areas of research involving interactions starting from the earth's interior to the magnetosphere. The lithosphere, atmosphere and ionosphere act critically during earthquake preparation period. The cause and effect of the precursory anomalies registered on these systems during pre-seismic time is difficult to be interpreted alone. Therefore, LAI coupling theory unifies the understanding of all the mechanisms together. The relative tectonic movements create stress. During the stress accumulation phase various geophysical precursors get registered spanning from earth degassing, metallic aerosols emanation, ground water uplift, geoelectric potential change, increase in P-wave velocity, anomalous vertical electric field and ionospheric precursors. Among these, radon plays a crucial role by ionizing the near-surface atmosphere. These ions coalesce with water molecules to decrease the relative humidity, increase in air temperature via changed air conductivity. The atmospheric conductivity gets altered due to ion cluster formation via radon emanation. This anomalous electric field further couples to alter the ionospheric parameters.

CONCLUSIONS

The present study analyzed twelve recent high-magnitude earthquakes ($M \geq 6.0$) that occurred between 2017 and 2024 to investigate ST as a potential thermal pre- and post- seismic activity. The results indicate that the normalized surface temperature anomaly, shows a consistent anomalous increase in nearly all of the examined events. For most earthquakes, these anomalies were observed within approximately 1-9 days prior to the mainshock. In addition, significant variations in normalized ST were detected during both the pre- and post-seismic periods, with pre- earthquake anomalies ranging from about 6.15% to 116.6% and post- earthquake anomalies ranging from approximately 17.69% to 83.95% above the background noise. The magnitude and temporal behavior of these anomalies vary among events, reflecting differences in crustal properties at the epicentral regions as well as atmospheric and meteorological conditions, such as humidity and water vapor content. The water vapor accumulates probably help trap the warmth at the surface and makes the ST

over the EPZ increase, which is consistent with the observed increase in ST prior to major earthquake occurrences.

Acknowledgments

The authors gratefully acknowledge the NCEP-NCAR for providing the Surface Temperature (ST) data. The authors also thank the United States Geological Survey (USGS) team site for sharing earthquake information.

Author Credit Statement

Problem was identified by Ananna Bardhan and data analysis was conducted by Pooja Sharma. All authors collaborated in writing the paper.

Data availability

The paper provides the data and website information.

Compliance with ethical standards

The authors declare no conflict of interest and confirm adherence to copyright norms.

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Received on: 16-09-2025; Revised on: 02-06-2026; Accepted on: 02-06-2026

Reanalysis dataset-driven assessment of super cyclonic storm Fani and extremely severe cyclonic storm Amphan over the Bay of Bengal, India

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ABSTRACT

This study uses European centre for medium-range weather forecasts (ECMWF) reanalysis 5 (ERA5) and Indian monsoon data assimilation and analysis (IMDAA) reanalysis datasets, to examine the characteristics and evolution of extremely severe cyclonic storm Fani and super cyclonic storm Amphan over the Bay of Bengal. Key parameters analysed include track, track error, mean sea level pressure, maximum sustained wind, potential vorticity and wind vector. For Amphan, all the parameters are assessed, while for the Fani, potential vorticity, and wind vector are not included. Results from both reanalysis datasets are compared with IMD observations. The comparison reveals that IMDAA provides better representation of most dynamical and structural features such as landfall-time track, mean sea level pressure, maximum sustained wind, potential vorticity, and wind fields particularly for Amphan. Conversely, ERA5 performs slightly better in estimating track errors for both cyclones. Overall, IMDAA dataset emerges as more reliable for analysing the tropical cyclones over the Bay of Bengal.

Keywords: Tropical cyclones, Fani, Amphan, Reanalysis datasets, Bay of Bengal.

INTRODUCTION

Tropical cyclones pose major threats to coastal regions due to strong winds, storm surges, and landfall impacts. About 80 tropical cyclones occur globally each year, with the North Indian Ocean contributing nearly 7%, including 1% over the Arabian Sea and 6% over the Bay of Bengal (McBride and Fraedrich, 1995; Kalnay et al., 1996). Tropical cyclone 'Fani' rapidly intensified into an extremely severe cyclonic storm before striking Odisha, causing widespread damage despite effective evacuations. One year later, 'Amphan' became the first super cyclone over the southeast Bay of Bengal in two decades, resulting in severe socio-economic impacts.

Recent studies using mesoscale models and data assimilation have improved understanding of tropical cyclone track, structure, and intensity (Montgomery et al., 2006; Rajasree et al., 2016). This study highlights the role of global reanalysis datasets, which are available at resolutions from 0.25° to 0.125°, and are essential for diagnosing tropical cyclone behaviour, conducting climatological analyses, and providing initial and boundary conditions for regional forecasting models (Thorne and Vose, 2010).

Extensive studies have examined the use of reanalysis datasets and numerical simulations for diagnosing tropical cyclone behaviour over the North Indian Ocean (Mohanty et al., 2004; 2010; Osuri et al., 2013). Reanalysis-based studies have contributed to tropical cyclone climatology, large-scale mechanisms of tropical cyclone evolution, power dissipation estimates, and analyses of tropical cyclone track, intensity, and life cycle (Maloney et al., 2000; Srivier and Huber, 2006; Schenkel and Hart, 2012). Metrics for representing outer wind structures in reanalysis datasets have also been refined

(Murakami, 2014; Bian et al., 2021). Despite forecasting improvements, further advancements in track and intensity prediction remain necessary (Mohapatra et al., 2013a, b; 2015)

Several recent studies have focused on 'Fani' and 'Amphan'. Alam (2020) assessed weather research forecast (WRF) parameterization sensitivities for simulating Fani, while Mohan et al. (2022) examined sea surface temperature-driven intensification mechanisms of both cyclones.

Naskar and Pattanaik (2023) linked cyclone intensity to environmental factors such as sea surface temperature, instability, moisture, wind shear, and latent heat flux and Hossain et al. (2022) evaluated weather research forecast model performance for the genesis and intensification of Fani.

Besides, Johari et al. (2023) analysed land surface model impacts on track, mean sea level pressure, maximum sustained wind, and rainfall and Kumar et al. (2024a) analysed that spatial distribution of wind and vertical integrated moisture transport using weather research forecast model are well matched with the ERA5 reanalysis datasets. Apart from these studies, Kumar et al. (2024b) simulated tropical cyclone 'Amphan' using weather research forecast model.

Present study compares two reanalysis datasets-ERA5 and IMDAA by evaluating key characteristics of landfalling tropical cyclones 'Fani' and 'Amphan' over the Bay of Bengal. The analysis focuses on track, track error, maximum sustained wind, mean sea level pressure, wind vectors, and potential vorticity. Additionally, the study relates potential vorticity fields to their mathematical formulations and examines synoptic features associated with both cyclones to improve understanding of their evolution and dynamics.

DATA AND METHODOLOGY

Data

The Indian monsoon data assimilation and analysis (IMDAA) is a regional atmospheric reanalysis datasets focusing on the Indian subcontinent. The IMDAA system is based on four-dimensional variational assimilation scheme and its Unified Model with a 6-hour intermittent data assimilation cycle. IMDAA uses horizontal grid resolution of 12 km ($0.12^\circ \times 0.12^\circ$) with 63 vertical levels reaching upwards to an approximate height 40 km. By integrating these diverse datasets, IMDAA endeavours to enhance the fidelity and reliability of its reanalysis, enabling a more accurate depiction of atmospheric conditions over the specified period (Ashrit et al., 2020). The Extended Kalman Filter-based land data assimilation system of the IMDAA is used to update the soil moisture initial condition in the model with sea surface temperature data set version 2 (Titchner and Rayner, 2014). In 2016, ECMWF replaced the highly successful ECMWF reanalysis (ERA)-Interim reanalysis with ERA5 reanalysis. Hersbach et al., (2018) discussed the ERA5 configuration and its compatibility with other reanalysis datasets. ERA5 dataset is used to show a cyclonic activity climatology for the Mediterranean region over the course of 40 years (1979- 2018) (Aragão and Porcù, 2022)

IMD observed datasets, including track, mean sea level pressure, and maximum sustained wind, are important for meteorological research. In this work, IMD observed datasets is used to analyze track, mean sea level pressure, and maximum sustained wind of tropical cyclone 'Amphan' and 'Fani' for validation of ERA5 and IMDAA reanalysis datasets. The horizontal resolution IMDAA and ERA5 is taken same for the analysis of track, track error, maximum sustained wind, and mean sea level pressure for both tropical cyclones, 'Fani' and 'Amphan'. On the other hand, the horizontal resolution IMDAA and ERA5 is taken originally available that is 12 km and 31 km for the analysis of wind distribution, potential vorticity, and vertical integrated moisture transport of tropical cyclone 'Amphan'.

Methodology

The root mean square error (RMSE) of the reanalysis datasets track, mean sea level pressure, and maximum sustained wind values have been computed using the following formula:

$$RMSE = \sqrt{\frac{\sum_{j=1}^n (x_j - y_j)^2}{n}} \quad (1)$$

where, j is a variable, n represents number of data points, x_j represents actual observational time series, and y_j represents reanalysis datasets time series.

The mean absolute percentage error (MAPE) of the reanalysis datasets mean sea level pressure, and maximum sustained wind values have been calculated using the following formula:

$$MAPE = \left(\frac{\sum_{j=1}^n \frac{|x_j - y_j|}{x_j} \times 100}{n} \right) \quad (2)$$

where, j is a variable, n represents number of data points, x_j represents actual observational time series, and y_j represents reanalysis datasets time series. 10 m wind is plotted using discrete values with time.

Potential vorticity is calculated directly on the model grid using assimilation system in ERA5 reanalysis datasets. For IMDAA, potential vorticity is not given directly and absolute and relative vorticity are also not provided on model levels. Therefore, potential vorticity is estimated directly from the reanalysis winds and temperatures (interpolated to isentropic surfaces) using a method similar to the ones described by Newman et al. (1989) and Manney and Zurek (1993). Potential vorticity on an isentropic surface is given by

$$\text{Potential Vorticity} = -g (\zeta_\theta + f) \partial \theta / \partial p \quad (3)$$

where g is gravity, f is the Coriolis parameter, p is pressure, θ is potential temperature, and ζ_θ is the component of relative vorticity orthogonal to the potential temperature (θ) surface. Specifically, for IMDAA, relative vorticity is derived from the zonal and meridional reanalysis winds.

RESULTS AND DISCUSSION

The following results are based on the datasets using IMDAA and ERA5 reanalysis datasets. Various parameters, including track, track error, maximum sustained wind, mean sea level pressure, spatial distribution of the wind field, and potential vorticity for 'Fani' and 'Amphan' are thoroughly examined. The accuracy of the best track results is assessed by comparing them with IMD observations.

Track and track error

The tracks of tropical cyclone 'Fani' using the 201-hours IMDAA reanalysis, ERA5 reanalysis, and India Meteorological Department (IMD) observed data, are presented in Figure 1(a) at a 31 km resolution. During the pre-landfall period both the IMDAA and ERA5 reanalysis datasets exhibited, reasonably accurate track and cyclone landfall position. The northward track of cyclonic storm 'Fani' is effectively captured by the ERA5 reanalysis datasets. The simulated paths for both reanalysis datasets closely match the observed track up to the landfall time, with the ERA5 reanalysis track showing greater proximity to the observed

track at the landfall point. Around the landfall period, both the IMDAA and ERA5 reanalysis datasets accurately tracked cyclonic storm 'Fani', but the ERA5 reanalysis slightly underestimated its proximity to the observed track at the landfall point. While the ERA5 reanalysis closely mirrored the observed path, the IMDAA reanalysis tended to overestimate the distance from observational points. Tropical cyclone 'Fani' made landfall on the Odisha coast of India between 0230 and 0430 UTC on May 3, 2019. The IMDAA and ERA5 reanalysis datasets indicated landfall times of 0400 and 0600 UTC on May 3, 2019, respectively, with no discernible time delay in the landfall for the ERA5 reanalysis datasets. The root mean square error for ERA5 reanalysis datasets is 0.62, while the root mean square error for IMDAA reanalysis datasets is 1.27. Therefore, the root mean square error is significantly lower for the ERA5 reanalysis datasets, compared to the IMDAA reanalysis datasets. The IMDAA reanalysis datasets captured the most deviated track, while the ERA5 reanalysis datasets-maintained a less deviated track throughout the cyclone period. Figure 1(b) depicts the observed tracks of super cyclonic storm 'Amphan'

using the 132 hours IMDAA and ERA5 reanalysis datasets at a 31 km, and compared with the observed track provided by the India Meteorological Department (IMD). The observed track spans from May 16, 2020, 0000 UTC, to May 21, 2020, 1200 UTC.

Both reanalysis datasets closely align with the observed track throughout the lifespan of the tropical cyclone. In comparison with the IMD track, the cyclone track analysis by both reanalysis datasets is more accurate and consistent when initialized. During deep depression and severe cyclonic storm, both reanalysis datasets exhibited higher track errors. During intense stages, both reanalysis datasets are well matched with low track error compared to initial stages. However, during the intense stages, the IMDAA reanalysis dataset's track closely paralleled the IMD observed track compared to the ERA5 reanalysis dataset. Notably, both reanalysis datasets exhibit close proximity to each other over the ocean. However, distinctions in track emerge during and post-landfall.

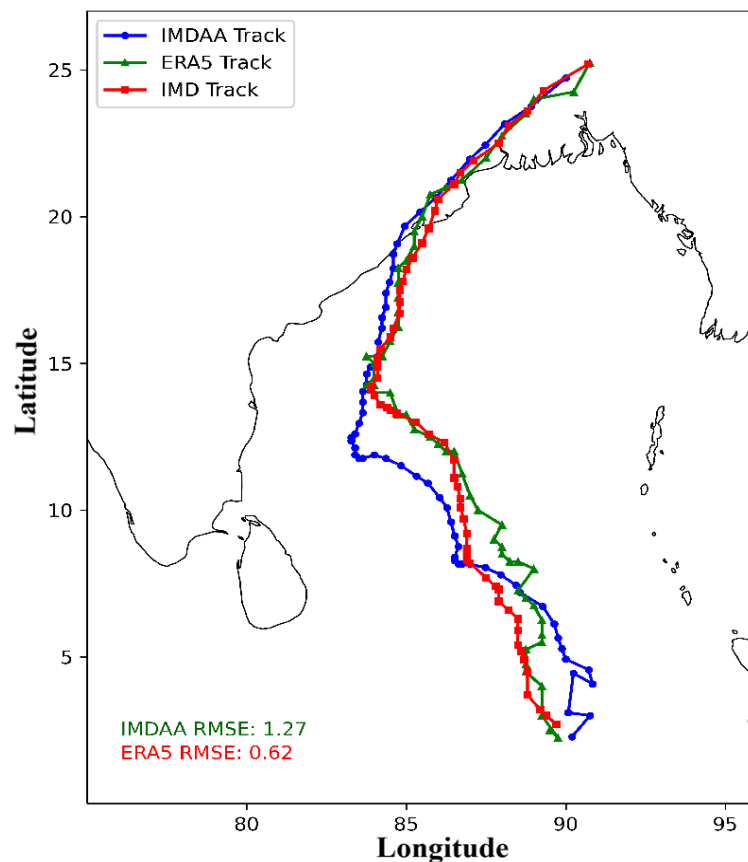


Figure 1a. Track of 'Fani' at various stages from pre- landfall (26 April, 2019) to post- landfall (05 May, 2019) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

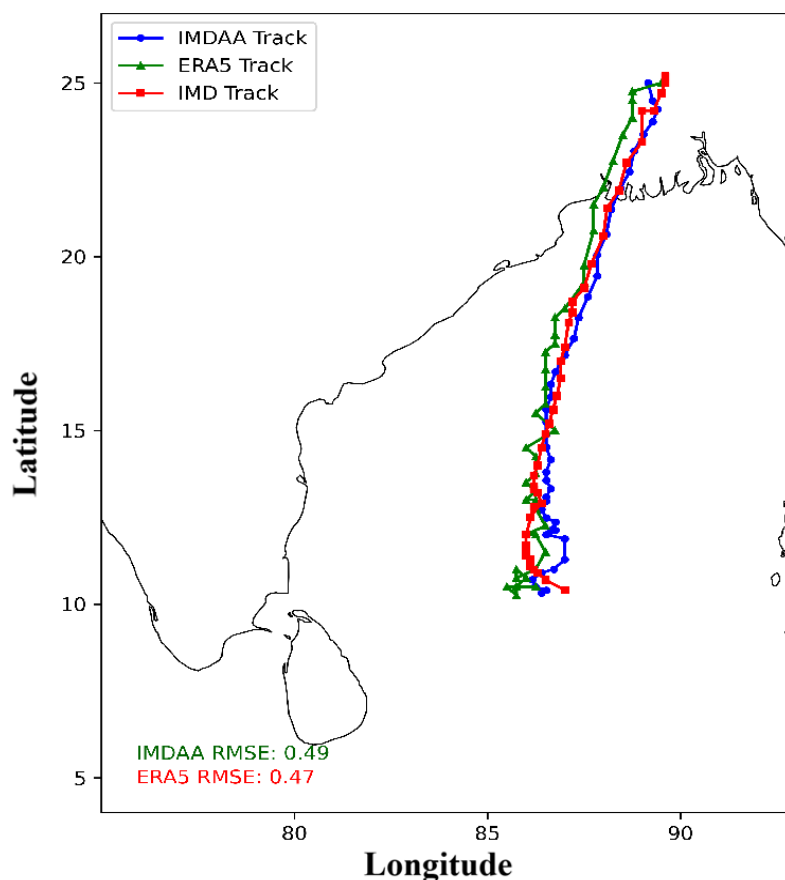


Figure 1b. Track of ‘Amphan’ at various stages from pre- landfall (16 May, 2020) to post- landfall (21 May, 2020) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

The root mean square error of track position is 0.49 and 0.47 using IMDAA and ERA5 reanalysis datasets, respectively, indicating relatively high precision. The position of ‘Amphan’ landfall in IMDAA reanalysis datasets almost matches the landfall position in IMD observed datasets, while the root mean square error of track position is less in ERA5 reanalysis datasets than in IMDAA datasets. Thus, during deep depression and severe cyclonic storm, both reanalyses tend to overestimate track errors, revealing challenges in accurately predicting these stages. However, during intense stages, the IMDAA reanalysis dataset closely parallels the observed track, showcasing its tendency to underestimate errors compared to ERA5, particularly noticeable during landfall where ERA5 exhibits higher errors in both position and time.

Mean sea level pressure

The Figures 2a, b, illustrate the temporal progression of mean sea level pressure for the ‘Fani’ and the ‘Amphan’. These figures provide a comparative analysis between the observed datasets from the India Meteorological Department (IMD) and the reanalysis datasets from IMDAA and ERA5. In Figure 2(a), during initial and intense stages, the ERA5 reanalysis datasets show the highest mean sea level pressure values than IMDAA

reanalysis datasets and the IMD observed datasets. Conversely, the IMDAA reanalysis datasets consistently exhibit the lowest mean sea level pressure values compared to ERA5 reanalysis datasets throughout this period. Specifically, from April 28 to April 30, IMDAA reanalysis datasets record lower mean sea level pressure values than both IMD observed datasets and ERA5 reanalysis datasets. A close alignment between IMD observed and IMDAA reanalysis datasets is observed during intense stages, while ERA5 exhibits consistently higher mean sea level pressure values throughout the same period. This suggests that during the intense stages of the cyclone, the IMDAA reanalysis datasets provide a more accurate representation of mean sea level pressure compared to both the IMD observed datasets and ERA5 reanalysis datasets.

The pressure drop in IMDAA reanalysis datasets is notably higher than that in IMD observed datasets around landfall. The lowest mean sea level pressure value captured by ERA5 reanalysis datasets is 973.38 hPa, higher than the lowest value of mean sea level pressure (940.78 hPa) captured by IMDAA reanalysis datasets throughout the period. The lowest mean sea level pressure value captured by IMDAA reanalysis datasets is closer to the IMD observed datasets mean sea level pressure

value (932 hPa) than the ERA5 reanalysis datasets mean sea level pressure value. The intensity of pressure drop is significantly higher in IMD observed datasets than in IMDAA and ERA5 reanalysis datasets before landfall, occurring from May 2, 2019, 0600 UTC, to May 2, 2019, 2100 UTC. Both IMDAA and ERA5 reanalysis datasets mean sea level pressure at the time of IMD observed least value of mean sea level pressure (932 hPa) were 947.58 and 974.24 hPa, respectively.

ERA5 overestimated with a root mean square error of 20.37 hPa and a mean absolute percentage error of 1.72% throughout the simulation time. In contrast, the IMDAA reanalysis datasets estimated with a root mean square error of 6.85 hPa and a mean absolute percentage error of 0.57% throughout the simulation time, with minor exceptions. Thus, ERA5 tends to overestimate the observed mean sea level pressure, as indicated by its higher root mean square error and mean absolute percentage error compared to IMDAA. Around the time of hitting the Odisha coast, ERA5 reanalysis datasets mean sea level pressure followed the pattern of observed mean sea level pressure, while IMDAA reanalysis datasets differed from the observed pattern, showing lower values than the observed ones. The deviation of mean sea level pressure between IMDAA reanalysis datasets and IMD observed datasets is much higher than ERA5 reanalysis datasets around the landfall of 'Amphan'. However, the mean sea level pressure is lower than ERA5 reanalysis datasets and IMD observed datasets. In Figure 2b,

ERA5 reanalysis datasets exhibit the lowest values of mean sea level pressure from May 16, 2020, 0000 UTC, to May 17, 2020, 1500 UTC, compared to both the IMD observed datasets and IMDAA reanalysis datasets. However, the IMDAA reanalysis datasets show the lowest mean sea level pressure values from May 19, 2020, 1500 UTC, to May 21, 2020, 0600 UTC.

During the initial deep depression to severe cyclonic storm, the IMD observed and IMDAA reanalysis datasets exhibit close agreement. However, thereafter, ERA5 consistently shows lower mean sea level pressure values from May 16 to May 19, but the pressure falls in both reanalysis datasets do not surpass those observed around landfall. Despite ERA5 lower mean sea level pressure values, the lowest value captured by ERA5 (942.59 hPa) is slightly higher than that of IMDAA (940.47 hPa) throughout the period. Additionally, the intensity of pressure falls is notably higher in IMD observed datasets compared to IMDAA and ERA5 reanalysis datasets during severe cyclonic storm to super cyclonic storm. Notably, during cyclone weakening, observed mean sea level pressure aligns closely with ERA5 reanalysis datasets, but during intensification, both reanalysis datasets mean sea level pressure values diverge significantly from the IMD observed datasets. At the time of IMD observed minimum mean sea level pressure (920 hPa), both reanalysis datasets (IMDAA and ERA5) report values of 956.72 and 955.18 hPa, respectively.

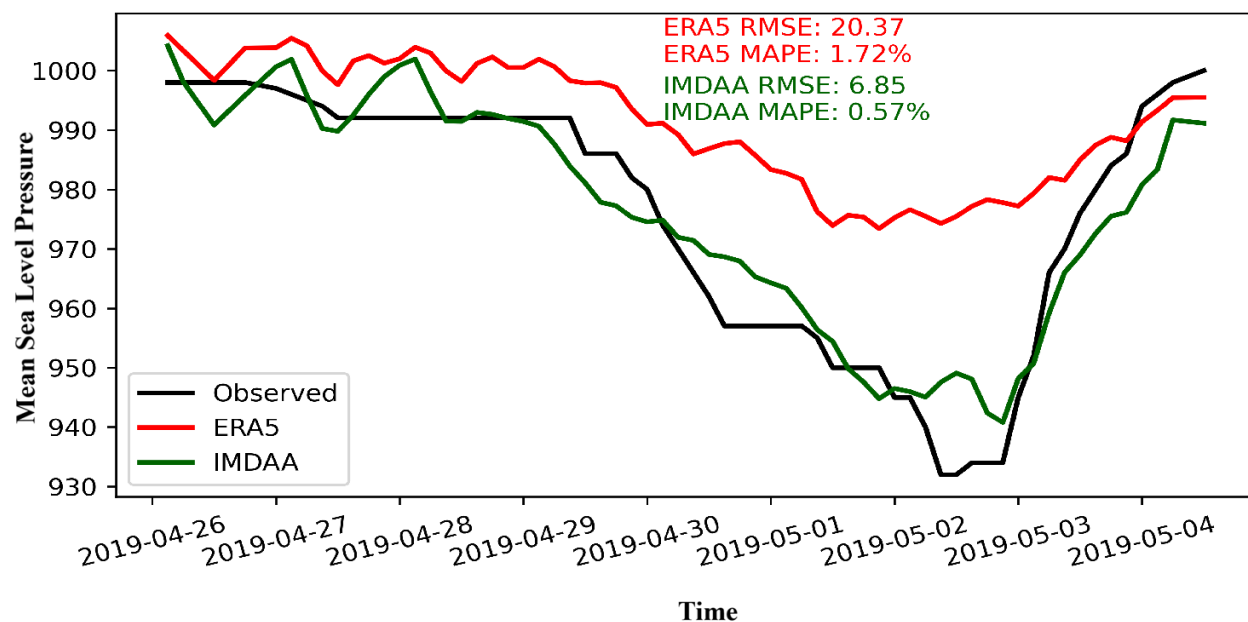


Figure 2a. Mean sea level pressure of 'Fani' at various stages from pre- landfall (26 April, 2019) to post- landfall (04 May, 2019) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

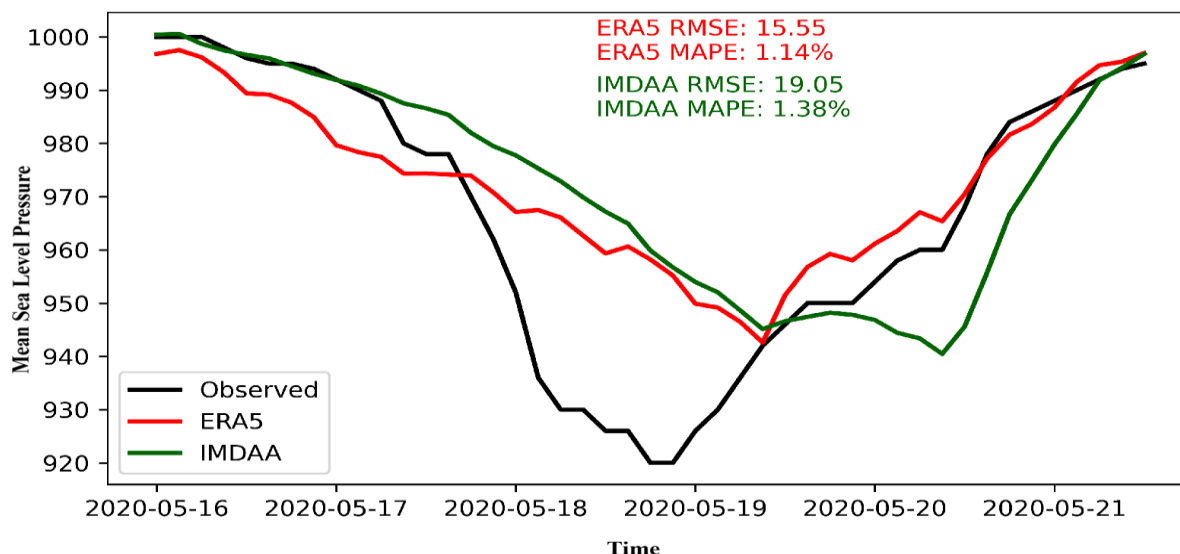


Figure 2b. Mean sea level pressure of ‘Amphan’ at various stages from pre- landfall (16 May, 2020) to post- landfall (21 May, 2020) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

The ERA5 reanalysis exhibits an overestimation, with a root mean square error of 15.55 hPa and a mean absolute percentage error of 1.14% across the entire simulation period, with only minor deviations. On the other hand, the IMDAA reanalysis datasets overestimated the observed values. The root mean square error and mean absolute percentage error of IMDAA reanalysis datasets are 19.05 hPa and 1.38% respectively. After landfall, ERA5 reanalysis datasets mean sea level pressure follows the pattern of observed mean sea level pressure, but IMDAA reanalysis datasets differ from the observed pattern. The ERA5 reanalysis datasets demonstrate a pressure fall intensity that closely aligns with the observed values up to the point of ‘Amphan’ making landfall. However, following the landfall of ‘Amphan’, there is a notable increase in the deviation of mean sea level pressure between the IMDAA reanalysis datasets and the observed values, surpassing that observed with ERA5.

Maximum sustained wind

The temporal variation of maximum sustained wind speed at 10 m height for ‘Fani’ and ‘Amphan’ from the India Meteorological Department (IMD) observed, IMDAA reanalysis, and ERA5 reanalysis datasets is presented in Figures 3(a, b), respectively. In Figures 3(a), it is observed that ERA5 reanalysis datasets consistently underestimate the maximum sustained wind, compared to IMDAA reanalysis datasets and IMD observed datasets during severe cyclonic storm to extremely severe cyclonic storm, while IMDAA tends to overestimate during very severe cyclonic storm to extremely

severe cyclonic storm up to severe cyclonic storm. During the intensification period, both reanalysis datasets capture lower values of maximum sustained wind than observed. However, around the landfall of ‘Fani’, IMDAA reanalysis captures a higher maximum sustained wind than observed. After May 3, 2019, 2100 UTC, the distribution of maximum sustained wind captured by ERA5 around landfall is almost similar to the IMD observed maximum sustained wind. At the time (1800 UTC on May 2nd) of the observed maximum sustained wind (115kt), the IMDAA and ERA5 reanalysis datasets maximum sustained wind values were 91.15 kt and 51.60 kt, respectively. It is evident that the maximum sustained wind captured by the IMDAA reanalysis datasets is lower than the observed maximum sustained wind at 10 m height, while ERA5 reanalysis datasets capture a much lower maximum sustained wind during the intensification. The IMDAA reanalysis datasets maximum sustained wind is almost similar to the IMD observed maximum sustained wind from May 3, 2019, 0900 UTC, to May 3, 2019, 1800 UTC. IMDAA reanalysis datasets capture a higher value of maximum sustained wind than ERA5 reanalysis datasets at most times but lower than the observed maximum sustained wind. The differences in both reanalysis datasets are high during the intensification period. However, the differences are low around the landfall of ‘Fani’. The root mean square error values for IMDAA and ERA5 reanalysis datasets are 15.44 and 33.75 hPa. The mean absolute percentage error values for IMDAA and ERA5 reanalysis datasets are 20.06% and 36.46%.

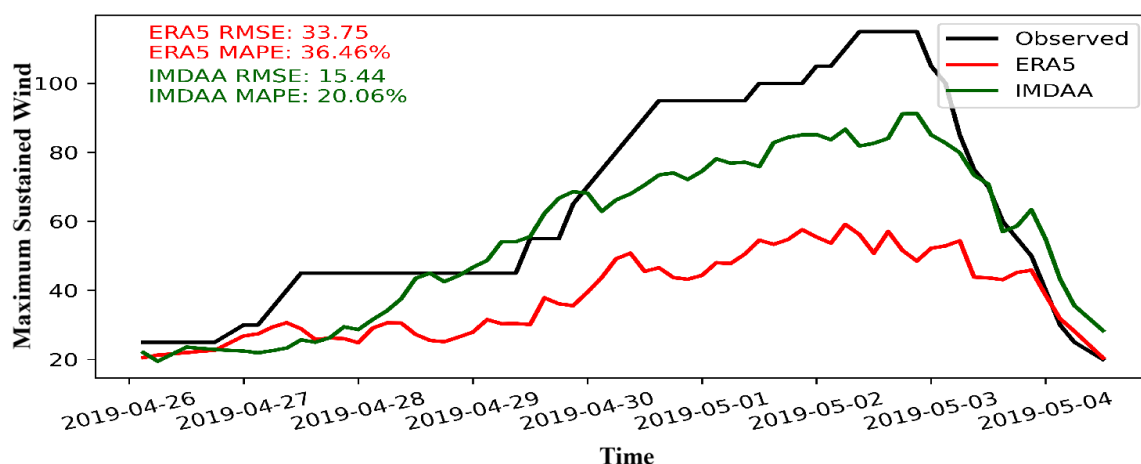


Figure 3a. Maximum sustained wind of 'Fani' at various stages from pre landfall (26 April, 2019) to post landfall (04 May, 2019) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

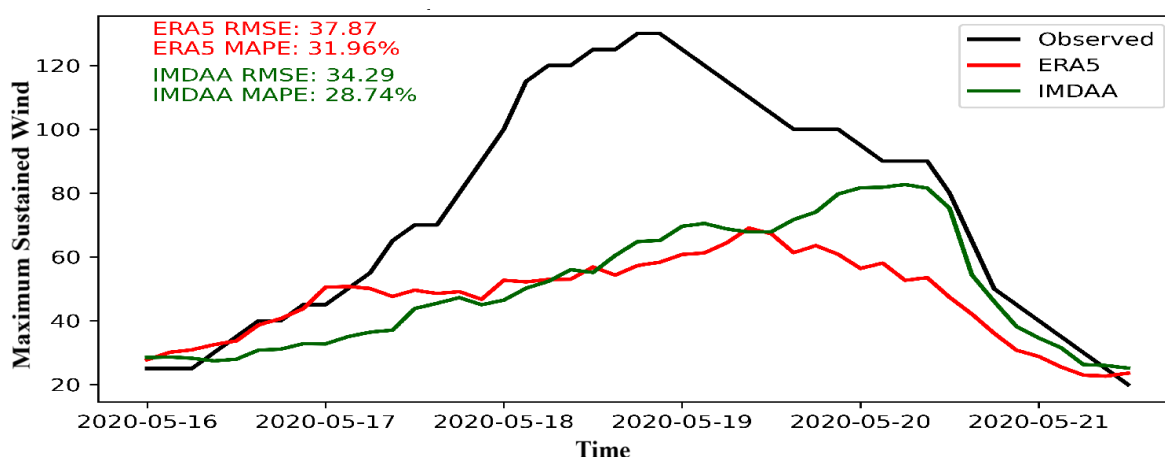


Figure 3b. Maximum sustained wind of 'Amphan' at various stages from pre- landfall (16 May, 2020) to post- landfall (21 May, 2020) using IMDAA, ERA5 reanalysis data sets and IMD observed data sets.

Figure 3(b) illustrates that ERA5 reanalysis datasets, consistently overestimate the maximum sustained wind compared to IMDAA reanalysis datasets during deep depression to super cyclonic storm formation, while IMDAA tends to underestimate during super cyclonic storm to depression formation, except around May 19, 2020, 0900 UTC. The maximum sustained wind at 10 m height captured by the IMDAA reanalysis datasets is lower than the observed maximum sustained wind from throughout the period. During the intensification period, both reanalysis datasets have captured a lower value of maximum sustained wind than observed. However, around the landfall of 'Amphan', IMDAA reanalysis captured higher maximum sustained wind than observed. After May 20, 2020, 0000 UTC, the distribution pattern of maximum sustained wind captured by IMDAA reanalysis datasets around landfall is closer to the IMD observed maximum sustained wind than ERA5 reanalysis datasets maximum sustained wind. Around the landfall period, the observed maximum sustained wind (130 kt), the IMDAA

and ERA5 reanalysis datasets maximum sustained wind values were 82.65 kt and 69.02 kt, respectively. It is evident that the maximum sustained wind captured by the IMDAA reanalysis datasets is lower than the observed maximum sustained wind at 10m height, while ERA5 reanalysis datasets captured a much lower maximum sustained wind during the intensification.

The ERA5 reanalysis datasets maximum sustained wind is almost similar to the IMD observed maximum sustained wind from May during depression to deep depression formation. IMDAA reanalysis datasets have captured a higher value of maximum sustained wind than ERA5 reanalysis datasets at most times, but lower than the observed maximum sustained wind. The differences in both reanalysis datasets and IMD observed datasets are high during the intensification period. However, the differences are low around the landfall of 'Amphan'. The root mean square error values for IMDAA and ERA5 reanalysis datasets are, 34.29 and 37.87 hPa. The mean absolute percentage error values for IMDAA and ERA5 reanalysis datasets are 28.74% and 34.29%.

Low level spatial distribution of winds

The low-level wind flow distribution (m/s) at 850 hPa for ‘Amphan’ is depicted in Figure 4, utilizing both IMDAA (Figure 4a-f) and ERA5 (Figure 4g-l) reanalysis datasets. In Figure 4, cyclonic wind flow at 850 hPa is observed in both reanalyses from May 16th to May 21st, 2020. The features exhibit a high level of correlation in both datasets. However, IMDAA captures an intense wind flow at an early stage (May 17th, 2020, 00 UTC) compared to ERA5 (May 18th, 2020, 00 UTC) and also depicts more intensification in wind flow than ERA5. IMDAA reanalysis (Figure 4a) illustrates a weak cyclonic circulation over southwest Bay of Bengal with a magnitude of 20-25 m/s. Subsequently, the cyclonic structure (Figure 4b-e) shifts more intensely towards northeast Bay of Bengal over time until May 20th, 2020 (00 UTC) with severe

wind velocities (> 50 m/s). Furthermore, the intense cyclonic structure is observed very close to the coastal area near Kolkata and eventually weakens (Figure 4f) on May 21st, 2020, near north Bangladesh with sustained wind velocities exceeding 30 m/s.

On the other hand, ERA5 reanalysis shows a feeble cyclonic circulation over southwest Bay of Bengal for an additional 24 hours (Figure 4g-h) with an almost identical magnitude of 20-25 m/s compared to IMDAA. The pattern of the cyclonic structure of wind velocity (Figure 4i-l) is almost similar to IMDAA reanalysis datasets. The cyclonic structure of strong wind flow is found to be similar in both reanalyses during the period from May 19th to May 21st, 2020 (Figure 4d-f and 4j-l) over the northeast sector.

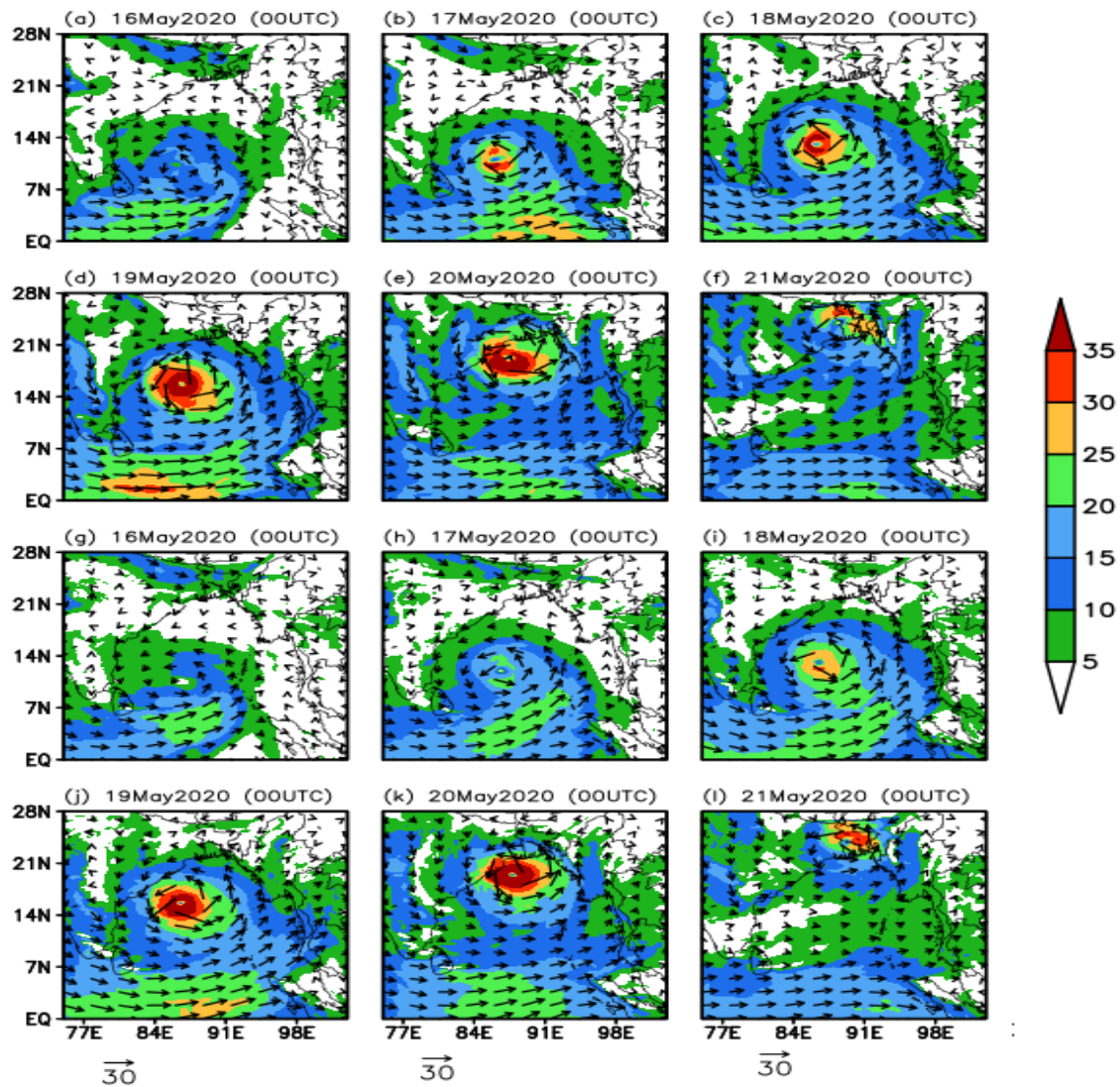


Figure 4. Distribution of low level (850 hPa) wind flow (m/s) at various stages from pre- landfall (16 May, 2020) to post- landfall (21 May, 2020) using IMDAA (a-f) and ERA5 (g-l) reanalysis data sets. The colour scale shows the range of wind speed.

Chan and Kwok (1999) observed that the wind speed is higher in the southwest sector of the storm during the initial stages compared to the other three sectors (northeast, northwest, and southeast) of the storm. As the storm's intensity increases, the wind gradually decreases in the sector compared to the other three sectors (northeast, northwest, and southeast). The strongest winds are observed in the eyewall of the storm in both reanalyses. It is noteworthy that the IMDAA reanalysis consistently captured well-organized cyclonic circulation throughout the entire period, outperforming the ERA5 reanalysis.

Potential vorticity analysis

The analysis of potential vorticity for 'Amphan' at 500 hPa, based on IMDAA and ERA5 reanalysis datasets during the

period from May 16th, 2020, to May 21st, 2020, at 00 UTC, is presented in Figure 5 (a-f) and Figure 5 (g-l), respectively. The intensity of potential vorticity increases from the initial phase to the developed phase around the inner region of 'Amphan'. The distribution of high potential vorticity values (greater than and equals to 3.0 PVU where, 1PVU equals to $10^{-6} \cdot \text{K} \cdot \text{m}^2 \cdot \text{kg}^{-1} \cdot \text{s}^{-1}$) is well captured by the IMDAA model in the inner region of 'Amphan'. The region of maximum potential vorticity shifts towards the northeast from the initial stage, and the peak potential vorticity region is observed near the coastal area of Bangladesh in both datasets (Figure 5 f, l). However, the intensity of potential vorticity in ERA5 is lower compared to IMDAA reanalysis datasets, attributed to the lower resolution in ERA5 reanalysis datasets.

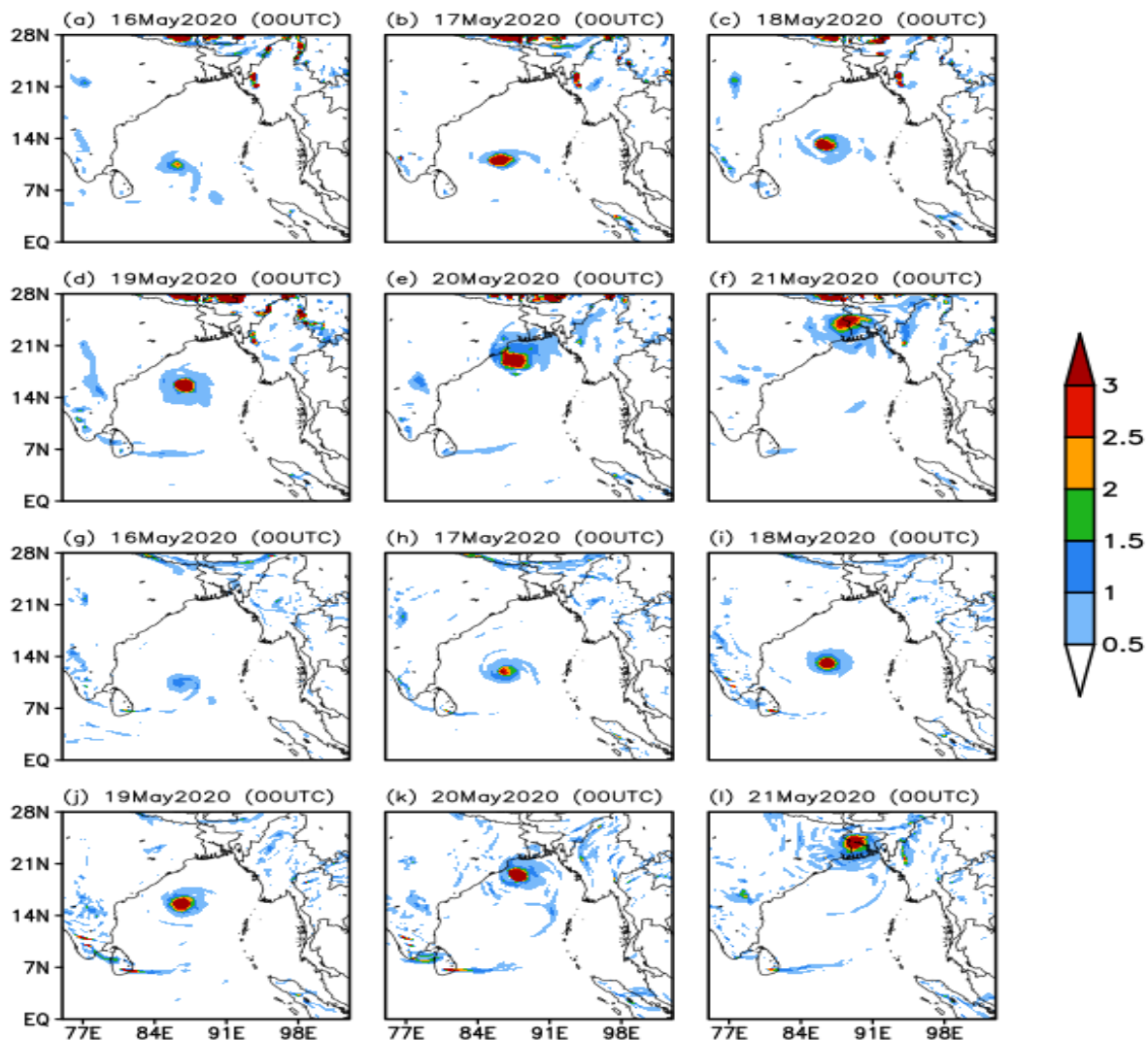


Figure 5. Potential vorticity of 'Amphan' at low level (500 hPa) at various stages from pre- landfall (16 May, 2020) to post- landfall (21 May, 2020) using IMDAA (a-f) and ERA5 (g-l) reanalysis data sets. The colour scale shows the range of potential vorticity.

Conclusions

This study evaluated the performance of IMDAA and ERA5 reanalysis datasets in representing key characteristics, track, track error, mean sea level pressure, maximum sustained wind, wind fields, and potential vorticity of ‘Fani’ and ‘Amphan’ over the Bay of Bengal. Overall, IMDAA demonstrated closer agreement with observations across most parameters not only due to its higher resolution and enhanced vertical structure but also at the same resolution of ERA5. For cyclone tracks, ERA5 provided better accuracy for Fani, showing lower root mean square error and better alignment with the observed track. In contrast, for Amphan, ERA5 showed positional and timing deviations during landfall, whereas IMDAA captured the landfall more accurately. In terms of intensity parameters, IMDAA outperformed ERA5 in capturing observed mean sea level pressure and maximum sustained wind for both cyclones. ERA5 consistently overestimated mean sea level pressure and underestimated maximum sustained wind, especially during intensification and landfall. Wind field analysis further showed stronger and better-organized circulation in IMDAA, particularly during early stages. Potential vorticity evolution indicated that IMDAA captured stronger inner-core structure and higher moisture transport compared to ERA5. Our study indicates that IMDAA provided a more accurate representation of structural and intensity-related features for both tropical cyclones, while ERA5 performed comparatively better only in certain track estimates.

Acknowledgement

The authors gratefully acknowledge the financial support given by the Earth System Science Organisation, Ministry of Earth Sciences, Government of India (Ref. No.-IITM/MM-III/2024/AI-P04/SO-004) to conduct this research under Monsoon Mission. The authors would like to thank Department of Mathematics, MMH College, Ghaziabad, UP, India, Department of Applied Mathematics, Gautam Buddha University, Greater Noida, UP, India to provide a good environment, library facility, and their technical support for this research. One of the author NK would like to acknowledge the support from IUCAA, Pune.

Author Credit Statement

Arun Kumar conceived and designed the study, conducted the data analysis and interpretation, and contributed to the critical review of the manuscript. Nitin Lohan helped in data analysis part. Sushil Kumar, Reena Nupur and Nagendra Kumar critically revised the manuscript for important intellectual content. Sushil Kumar reviewed and approved the final version of the manuscript for submission.

Data Availability

The IMD observed, ERA reanalysis, and IMDAA reanalysis data used in this study are available at https://rsmcnewdelhi.imd.gov.in/report.php?internet_menu=MzM=, <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?tab=download>, and <https://rds.ncmrwf.gov.in/datasets>.

Compliance with Ethical Standards

The authors declare that they have no conflicts of interest and that they have adhered to all applicable copyright and ethical guidelines.

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Received on: 18-07-2025; Revised on: 11-06-2026; Accepted on: 23-06-2026`

Comparative study of effective permeability and porosity at different gas hydrate bearing areas identified from NGHP-02 expedition in KG basin, eastern Indian offshore

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ABSTRACT

India's National Gas Hydrate Program (NGHP) Expedition-02 identified three areas, namely, B, C, and E in the Krishna Godavari (KG) offshore basin, where pressure core analysis and logging -while-drilling data, show that a significant amount of gas hydrate is present at several drilling locations in the designated areas. Since, permeability directly regulates the flow of gas and water from the reservoir, that influences both production efficiency and economic viability, estimating permeability is essential to understand the behaviour of gas production in gas hydrate reservoirs. At first, we construct an empirical permeability relation from multiple existing permeability models using downhole data calibrating with available core measurements that best represents the areas B, C, and E. In Area B, overall absolute permeability is about 1000 mD, whereas, effective permeability is about 100 mD. In Area C, overall absolute permeability is about 0.1 mD and effective permeability is about 0.01 mD. Whereas, in Area E, overall absolute permeability is about 0.01 mD and effective permeability is about 0.001 mD. According to our study, generally, a total porosity of about 60% is reduced to 50% as an effective porosity in Area B, whereas, a porosity of 40% is reduced to 10% as an effective porosity in Area C and E, due to high percentage of clay of approximately 65%. Our study predicts that a sediment with gas hydrate saturation of 80% of total porosity, has an effective permeability of 2% of its absolute permeability. We provide a relation of the effective permeability with total porosity and measured resistivity of each area, because such relations are particularly valuable when core samples are unavailable or disturbed during retrieval. Moreover, linking the resistivity with the effective permeability help to identify the gas hydrate morphology and predict gas production capacity for long-term production strategy.

Keywords: SNGHP-02, KG Basin (India), Effective permeability, Effective porosity, Gas hydrate saturation.

INTRODUCTION

High-permeability fault-fracture systems, allow high-flux methane to flow to shallow formations, where it forms gas hydrate in the stability zone (Riedel et al., 2010; Davies et al., 2014; Ren et al., 2023). In commercial energy recovery scenarios, accurate permeability estimation is crucial for understanding fluid dynamics, simulating gas hydrate dissociation, and assessing the potential for methane production (Konno et al., 2017; Wang et al., 2023; Ning et al., 2023). Direct techniques like formation tests and core measurements typically offer the accurate permeability. Nevertheless, these approaches are constrained by their high expense and restricted coverage. As a result, it has become more common to estimate permeability throughout the wellbore using empirical relations derived from the well log data.

The majority of the existing researches on hydrate reservoir, permeability estimation rely on empirical formulas. These models are sensitive to the input parameters particularly in heterogeneous and low porosity permeability reservoirs, where porosity-permeability relationships show notable scatter due to variations in pore throat curvature and shape, (Prakoso et al., 2017; Cai et al., 2019; Wu et al., 2023). Moreover, high clay content in sediments reduces effective porosity and disrupts pore connectivity, which directly impacts permeability (Mesri and Olson, 1971; Katsube and Connell, 1998). As a result, adding clay volume to porosity-permeability relations increases the accuracy of predictions, especially in formations

where sandy and muddy lithologies are interbedded (Mabiala et al., 2025; Pang et al., 2025). In a study, Dai and Seol (2014) presented a model based on simulations in a pore network model to investigate variation of permeability with distribution of pore sizes and heterogeneous hydrate distribution in the case of pore-filling hydrate. This suggests that the pore-filling habit, in which hydrate occupies the centres of pores, has a significant impact on permeability. Priest et al. (2015) measured the hydrate-bearing sediments' in-situ permeability using permeability tests on pressure cores. Additionally, sophisticated well-logging techniques like nuclear magnetic resonance (NMR) logging, among others, offer crucial inputs to predict permeability from well-known models such as Schlumberger Doll Research (Kleinberg et al., 2003; Zhang and Weller, 2021).

The purpose of this study is to examine all well-established permeability models and select a suitable one or derive a best suitable model for the gas hydrate reservoir of the Krishna-Godavari (KG) basin. Therefore, we show a method to predict permeability using the downhole data of wells and core measurements of permeability from the NGHP-02 expedition in the KG basin. Next, we use a well-established rock physics theory at each selected well to determine the gas hydrate saturation from sonic velocity, which is required to estimate the relative permeability at each well. Finally, the study offers a relationship between each area's observed resistivity, porosity and effective permeability, as these relationships are especially useful when core samples are missing or disrupted during retrieval.

STUDY AREA

Numerous studies related to the exploration and exploitation of natural resources have been conducted on the Krishna Godavari (KG) basin, formed during the Late Jurassic rifting of the Indian Plate from Gondwanaland, where the basement faulting created horst-and-graben structures enable the accumulation of over 5 km of sediments (Sastri et al., 1981; Rao, 2001; Yamamoto et al., 2018). The deep-water depositional systems including graded fans, channel-levee complexes, and mass-transport deposits define the basin. The cohabitation of organic-rich and coarse-grained sediments, increases the likelihood of gas hydrate formation, where the gas is primarily microbiological in origin (Lorenson and Collett, 2018). The Krishna and Godavari rivers provide the majority of the basin's sediment supply, with the Ganges–Brahmaputra system also contributing through the Bengal Fan (Alam et al., 2003; Rao et al., 2015). The NGHP Expedition 02 identified extensive sand-rich depositional systems by drilling, coring, and logging at 25 sites throughout the Areas B, C, and E (Collett et al., 2019). These coarse-grained sediments ranging in size from silts to gravels, offer ideal porosity and permeability for gas migration and hydrate formation as fracture-filling, or pore-filling or both. However, for all three

areas (B, C and E), primarily we select those sites having the core results for permeability, then the further sites we select that are nearest to the cored sites.

Twelve logging-while-drilling (LWD) sites were drilled inside a buried anticline structure in Area B, which is located at a water depth of approximately 2550 meters. These sites were classified into eastern and western transects (Figure 1), with key sites NGHP-02-16, 02-17, 02-20, 02-23, and 02-24 lying on the western transects are taken into consideration in this study. The hydrate accumulation in this area is facilitated by the sediments, which are primarily Late Miocene to Pliocene in age and are deposited in channel-levee systems (Raju, 2007; DGH, 2017; Dixit et al., 2019). However, Area C comprises of six sites targeting channel-levee systems along the mid- to lower continental slope. Within channel-fill deposits above the BSR, the sites NGHP-02-08 and NGHP-02-09, exhibit gas hydrate saturation in coarse sands and gravels. These sediments, which were generated in deep-water slope fan systems, are challenging to connect due to their complex lithofacies and younger age (post-Late Miocene to Pliocene). On the other hand, Area E consists of four sites, of which the present study selects NGHP-02-01 and NGHP-02-04 sites for investigation.

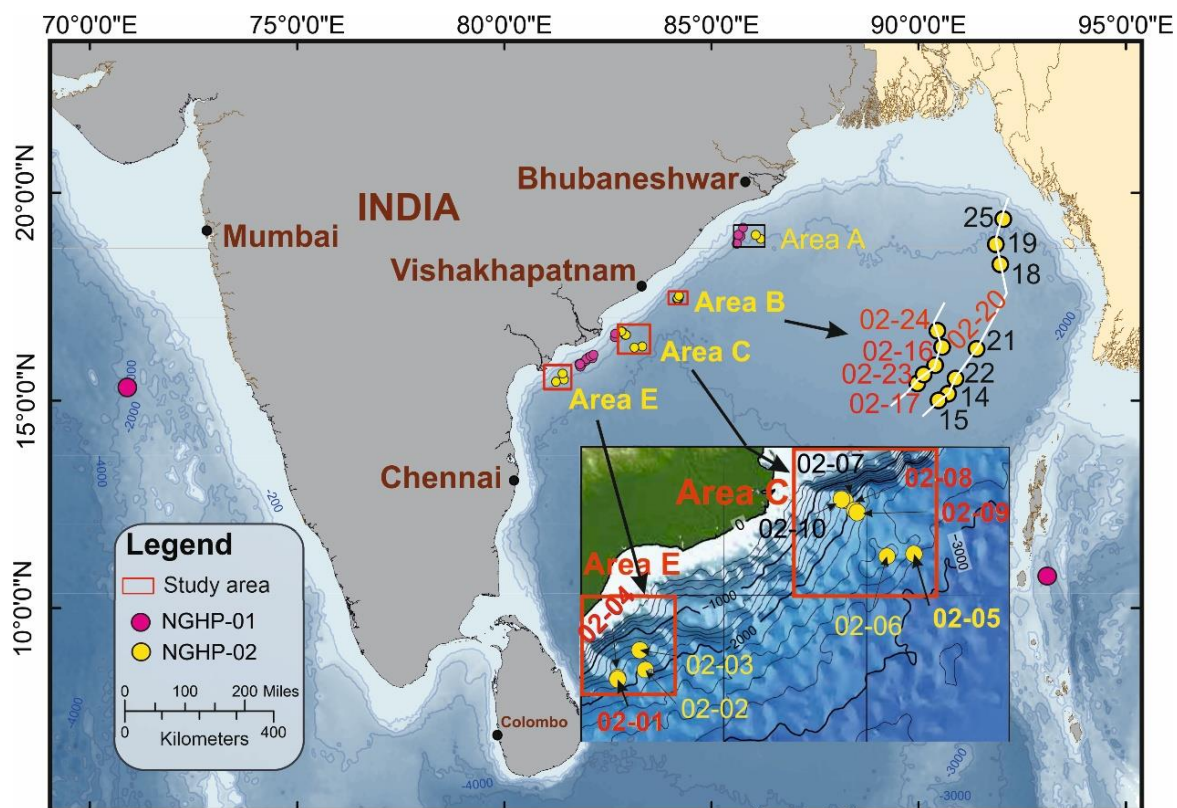


Figure 1. Drilling sites of NGHP-01 (pink circles) and -02 (yellow circles) expeditions. Locations of Area B, C, and E of NGHP-02 expedition in the Krishna Godavari Basin (Modified after Ghosh and Ojha, 2021). The wells numbered in red from Areas B, C, and E of NGHP-02 expedition are selected for this study.

DATA

This study utilizes logging-while-drilling (LWD) data from nine wells, including NGHP-02-17A, -23A, -16A, -24A, and -20A from Area B, NGHP-02-08A and -09A from Area C, and NGHP-02-01A and -04A from Area E. The study uses density log to estimate the total porosity while resistivity log provides direct advanced interpretations. Measurements on pressure cores collected from with or without gas hydrate zones from some wells are used for calibration. Variations in sediment physical properties among units, together with the site-specific differences in burial depth and unit thickness, lead to spatially variable effective reservoir quality and methane availability.

METHODOLOGY AND RESULTS

Permeability model

In this study, we aim to derive a permeability model for each of the three areas (B, C and E) of the KG basin by relating the absolute permeability (K_{ab}) with the effective porosity (ϕ_e) and irreducible water saturation (S_{wir}) that represent better each area. Therefore, we first examine the existing empirical relations (Tixier, 1949; Wyllie and Rose, 1950; Morris and Biggs, 1967; Timur 1968; Coates and Denoo, 1981) for the estimation of permeability using a methodical iterative approach.

Existing models for absolute permeability

Both theoretically and experimentally, correlations between permeability and more easily quantifiable metrics like effective porosity (ϕ_e) and irreducible water saturation (S_{wir}), have been thoroughly investigated. Among these are the models put forth by Tixier (1949), Wyllie and Rose (1950), Morris and Biggs (1967), Timur (1968), and Coates and Denoo (1981), which relate permeability to porosity and irreducible water saturation using regression-based techniques with different coefficients. Using ϕ_e and S_{wir} as the primary inputs, Tixier (1949) developed one of the first empirical correlations to predict the permeability from the well log data as shown in Equation 1.

$$K_{ab} = \left(250 \frac{\phi_e^3}{S_{wir}}\right)^2 \quad (1)$$

The equation lays the foundation for subsequent log-based prediction by offering a useful way for estimating permeability in the absence of core data. Wyllie & Rose (1950), who replaced a given surface area with S_{wir} , a parameter that is challenging to calculate directly, proposed an empirical variation of the Carman-Kozeny equation (Carman, 1937) shown in Equation 2. S_{wir} is a useful proxy for evaluating permeability from well log data since it is strongly correlated with the pore size.

$$K_{ab} = P \frac{\phi_e^Q}{S_{wir}^R} \quad (2)$$

Fitting the estimated results with the core data yields the constants P, R, and Q of Equation 2, which are determined by Wyllie and Rose (1950) as shown below in Equation 3,

$$K_{ab} = \left(100 \frac{\phi_e^{2.25}}{S_{wir}}\right)^2 \quad (3)$$

In order to estimate permeability from well log-derived parameters, namely ϕ_e and S_{wir} , Morris and Biggs (1967) proposed an empirical equation (Equation 4) that is more suitable for sandstone reservoirs.

$$K_{ab} = \left(80 \frac{\phi_e^3}{S_{wir}}\right)^2 \quad (4)$$

With an emphasis on formations with a clay content, Timur (1968) developed an empirical equation (Equation 5) to predict the permeability from well log data by integrating ϕ_e and S_{wir} .

$$K_{ab} = \left\{0.136 \left(\frac{\phi_e^{4.4}}{S_{wir}^2}\right)\right\} 10^5 \quad (5)$$

A popular empirical model for calculating permeability from well logs was also developed by Coates and Denoo (1981). It defines permeability as a function of ϕ_e and the ratio of irreducible water saturation to free fluid index, which is represented in Equation 6 as shown below,

$$K_{ab} = \left(100 \frac{\phi_e^{2(1-S_{wir})}}{S_{wir}}\right)^2 \quad (6)$$

Calculation of various input parameters for permeability modelling

In this part, we demonstrate how we use the observed log data to determine the total porosity, effective porosity, and irreducible water saturation.

Total porosity

Total porosity is the measure of all void spaces in the rock, regardless of their contribution to fluid flow. Density logs are the most accurate way to estimate total porosity among well logs since they are less impacted by gas and shale than neutron or sonic logs. Total porosity with depth is thus obtained using the following relation,

$$\phi_d = \frac{\rho_m - \rho_b}{\rho_m - \rho_f} \quad (7)$$

where ρ_b , ρ_f , and ρ_m are the sediment's bulk, fluid, and matrix densities, respectively.

Clay/or Shale Volume

In order to accurately characterise a reservoir, it is imperative to comprehend how permeability varies with lithology specially, clay volume. The best initial technique for determining shaliness is still to directly estimate clay volume from gamma ray logs. Equation 8 shown below illustrates the Larionov equation (Larionov, 1969), a popular empirical

relation for calculating clay volume (shale volume) in sedimentary rocks of Tertiary age based on gamma ray log data

$$V_{\text{clay}} = 0.083 * (2^{3.7 \cdot \text{IGR}} - 1) \quad (8)$$

Where, the gamma ray index IGR is expressed as below

$$\text{IGR} = \frac{\text{GR}_{\log} - \text{GR}_{\min}}{\text{GR}_{\max} - \text{GR}_{\min}} \quad (9)$$

Here, $\text{GR}_{\max}$, $\text{GR}_{\min}$ and $\text{GR}_{\log}$ in Equation 9 denote respectively the minimum, maximum, and log read values of the gamma ray.

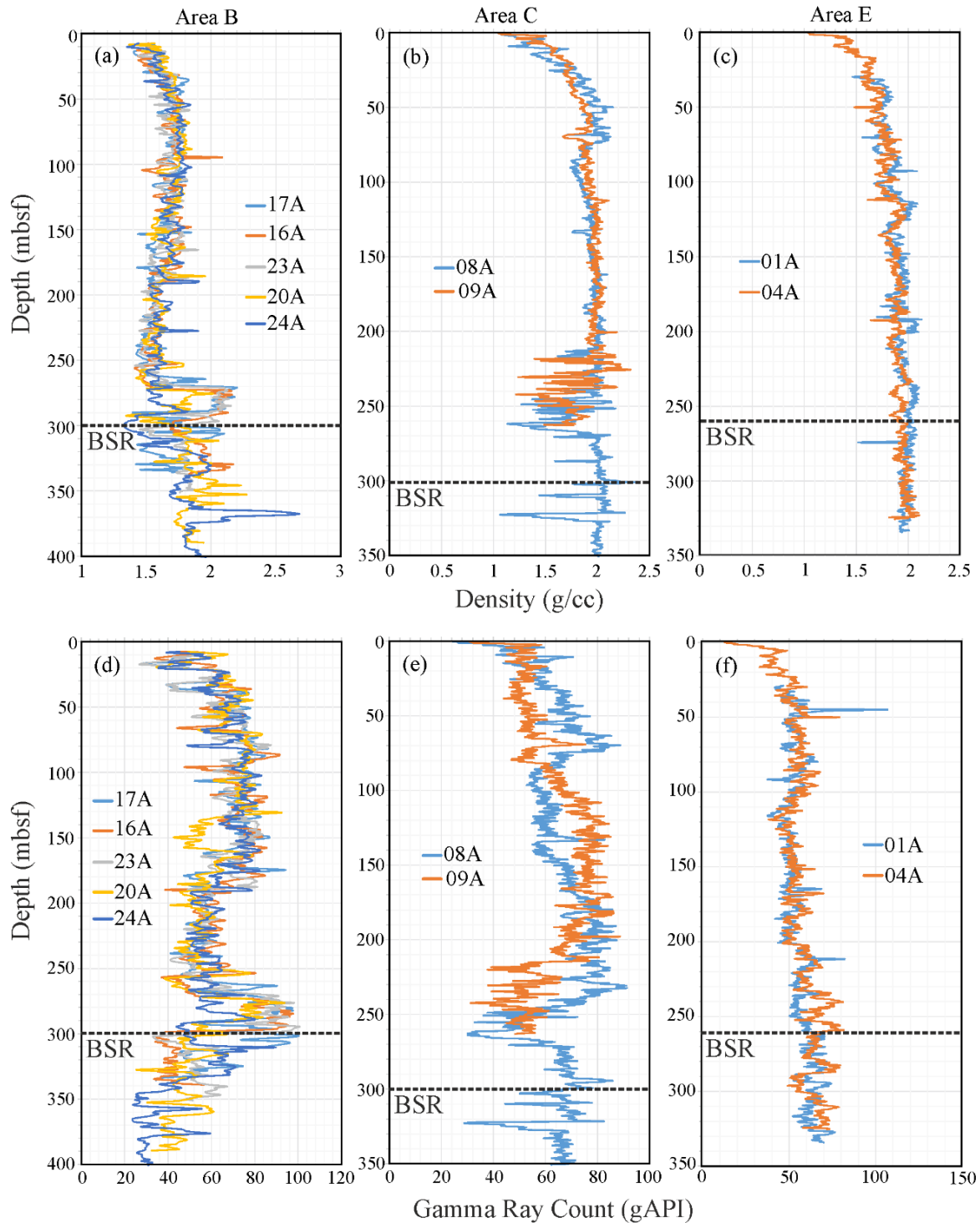


Figure 2. Measured density (a, b, c) and gamma ray (d, e, f) logs at selected wells in areas B, C, and E.

Effective porosity

The portion of total porosity that is interconnected is called effective porosity, which excludes the isolated pores or pore spaces filled by clay-bound water. To account for the presence of clay minerals, porosity estimated from the well logs usually has to be corrected. The adjusted measure, the effective porosity (ϕ_e), is usually lower than the total porosity (here, density porosity). Effective porosity is estimated using the shale correction model (Asquith and Krygowski, 2004) that is provided in Equation 10,

$$\phi_e = \phi_d * (1 - V_{\text{clay}}) \quad (10)$$

This equation assumes that the total porosity includes pore space that is occupied or affected by clay-bound water.

Irreducible water saturation

Buckles (1965) first proposed the idea that porosity and irreducible water saturation in carbonates and sandstones are hyperbolically connected. A number of researchers, including Morris and Biggs (1967), Holmes et al. (2009), later supported the idea shown below in Equation 11,

$$\phi_e * S_{\text{wir}} = \text{constant} \quad (11)$$

According to Konno et al. (2018), $S_{\text{wir}} = \sim 0.27$ for the site NGHP-02-16-A of the KG basin. By using the trial-and-error method, we find the value of constant in Equation 11 as 0.06 for the value of $S_{\text{wir}} = 0.27$. Equation 11 can therefore be expressed as follows,

$$S_{\text{wir}} = 0.06 / \phi_e \quad (12)$$

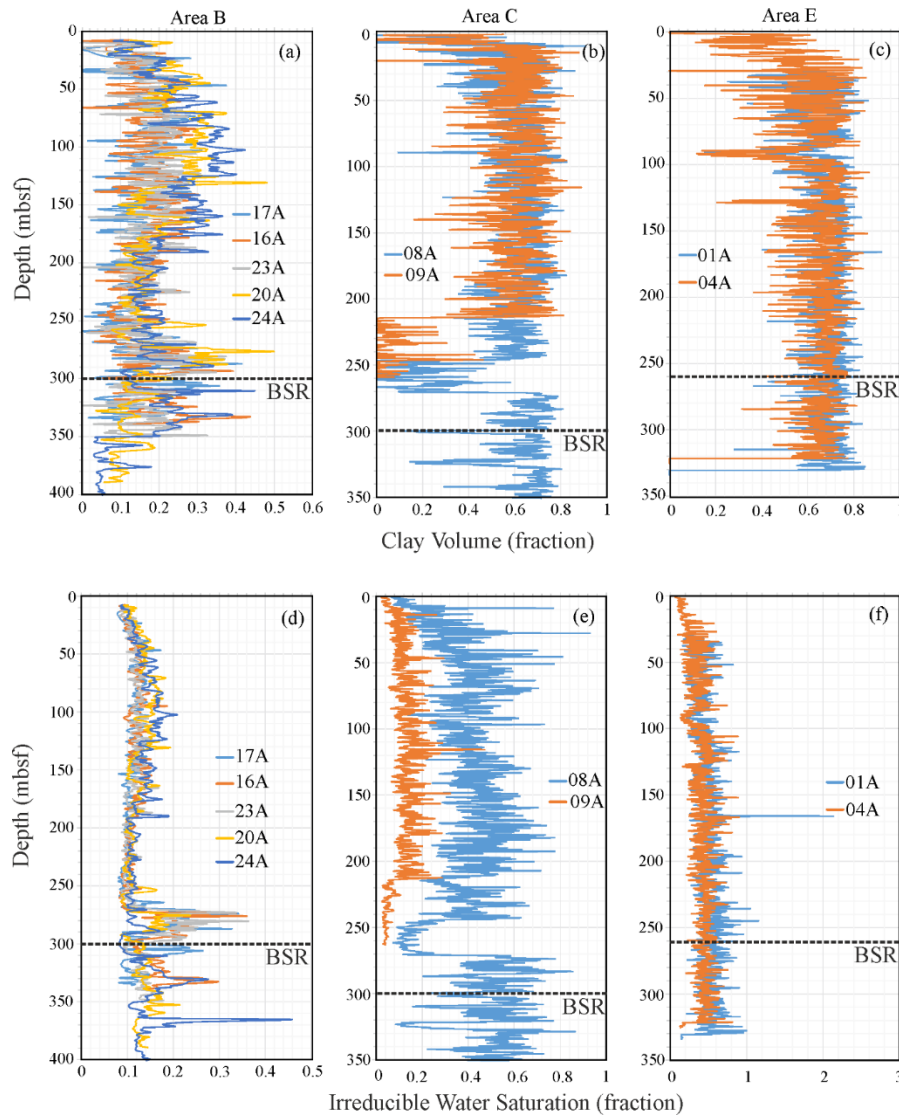


Figure 3. Calculated clay volume (a, b, c) and irreducible water saturation, S_{wir} , (d, e, f) logs at selected wells in areas B, C, and E.

Here, Equation (12) employs a Buckles number of 0.06, which was determined through calibration based on the reference irreducible water saturation ($S_{wir}=0.27$) reported for Area B by Konno et al. (2018). This value is also consistent with the Buckles number commonly adopted for fine-grained sandstones (Crain, 2018). The value of irreducible water saturation in other wells is assumed to be the same as that of site NGHP-02-16-A in our study due to the unavailability of site-specific S_{wir} data for all wells. In clay-rich intervals where the effective porosity was ≤ 0.06 , Equation (12) produced calculated S_{wir} values greater than 1.0. Since irreducible water saturation is physically constrained to values between 0 and 1, all calculated values exceeding 1.0 were excluded from further analysis and interpretation.

In this work, we first estimate the total porosity (ϕ_d) from the bulk density log (ρ_b) using Equation 7 for all wells under investigation, namely, NGHP-02-17A, -23A, -02-16A, -20A and -24A of Area B, NGHP-02-08A and -09A of Area C, and NGHP-02-01A and -04A of Area E in the KG basin. We use V_{clay} from the available spectrolith data of all wells, however, Equation 8 can be used to calculate the volume fraction of clay (V_{clay}) from the observed gamma log (Equation 9) in absence of the spectrolith data. Figure 2 shows the density and gamma-ray log data of the selected wells from areas B, C and E.

The calculated V_{clay} from the spectrolith data and irreducible water saturation are displayed in Figure 3. Gamma-ray logs in Figure 2 and V_{clay} calculated from spectrolith data in Figure 3 shows good trends with the data.

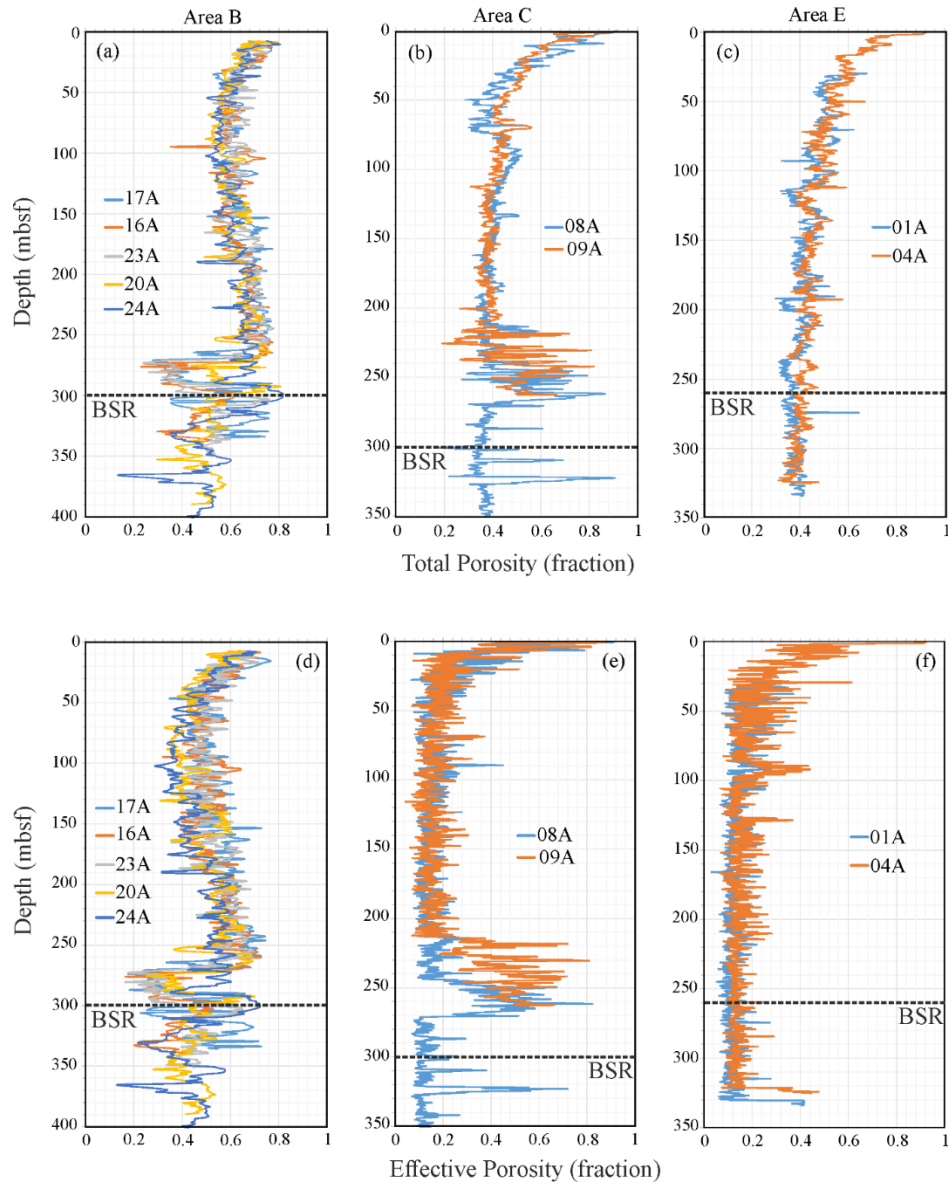


Figure 4. Calculated total porosity (a, b, c) and effective porosity (d, e, f) logs at selected wells in areas B, C, and E.

Now, Equation 10 provides the effective porosity using V_{clay} and Equation 12 provides the irreducible water saturation as discussed above. Here, the fluid-density (ρ_f) and matrix-density (ρ_m) of the sediment at all wells are respectively assumed as 1 g/cc and 2.64 g/cc. Figure 4 shows the calculated total porosity from the density log (Equation 7) and effective porosity (Equation 10) for comparison. We observe that the irreducible water saturation (Figure 3) increases and effective porosity (Figure 4) decreases with the increase of clay volume.

We expect three different porosity-permeability relations for the regions B, C and E of the KG basin. For this, we first derive a relation between the porosity and permeability (Equation 13) for Area B (Begum et al., 2026) through iterative parameter adjustments and calibrations to the existing models using the well log data and tuning with the core data available at wells NGHP-02-17A, -23A and -16A.

$$K_{ab} = 1200 * \phi_e^{6.6} * \left(\frac{1-S_{wir}}{S_{wir}}\right)^2 \quad (13)$$

However, we observe that the available core data from NGHP-02-09A of Area C, and NGHP-02-01A of Area E, match well with the derived permeability equation. Therefore, we consider Equation 13 represents best the KG basin that can be used to determine the absolute permeability from well log data in a comparable geological setting. Figure 5 displays the estimated permeability in the selected three areas (B, C and E) of the KG basin that shows good match with the available measurements from the core data.

Effective permeability modelling

Effective permeability is the term used to describe the decreased intrinsic permeability of sediments caused by the presence of gas hydrates. The following formulation of the effective permeability (K_{eff}), proposed by Masuda et al. (1997, 1999), is frequently used to represent the permeability reduction brought on by hydrate saturation:

$$K_{eff} = K_{ab} * (1 - S_h)^N \quad (14)$$

where, N represents the permeability reduction exponent and K_{ab} is the absolute permeability of the hydrate-free sediment. This formula explains how effective permeability decreases when hydrate saturation (S_h) increases. Theoretically, effective permeability equals the absolute permeability in the absence of hydrates and becomes zero when hydrates fully occupy the pore space. There is considerable uncertainty and variation in determining the exact value of N , as highlighted in several previous studies. However, this functional form consistently demonstrates good agreement with laboratory investigations that link hydrate saturation to permeability reduction, despite these differences (Masuda et al., 1997, 1999; Kumar et al., 2010; Chen et al., 2016). A key advantage of this model is its flexibility, allowing the parameter N to be calibrated to account for the specific pore habit of hydrate-bearing sediments or changes in pore structure with reservoir depth.

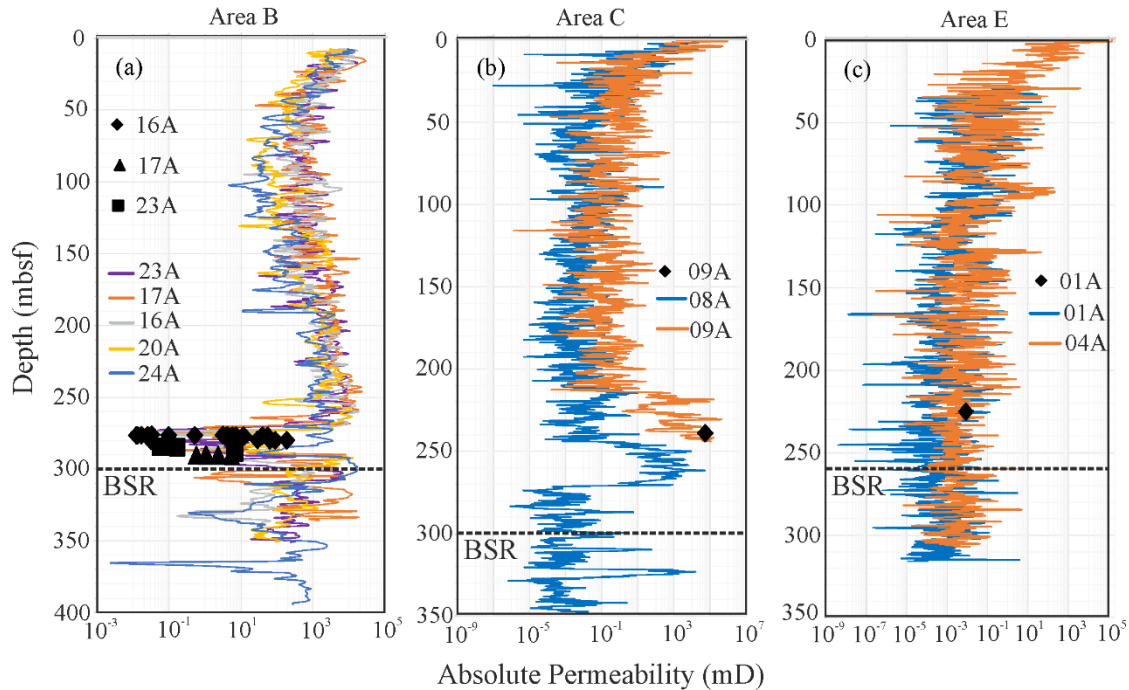


Figure 5. Calculated absolute permeability logs at selected wells in Area B (a), Area C (b), and Area E (c). Solid diamonds, triangles and squares denote the core measurements respectively for wells 16A, 17A, and 23 A.

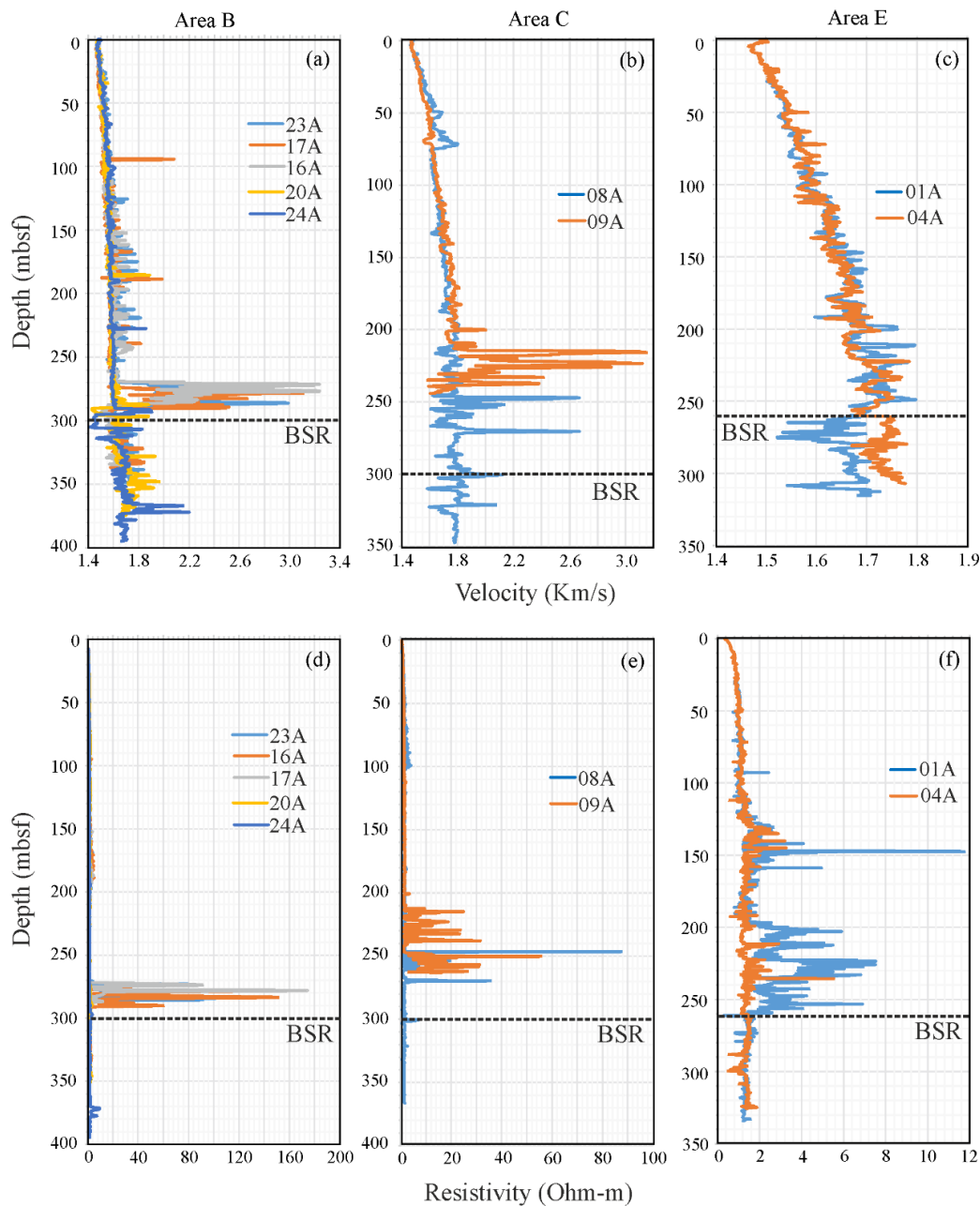


Figure 6. Measured sonic velocity (a, b, c) and resistivity (d, e, f) logs at selected wells in areas B, C, and E.

To account for gas hydrate morphology, Masuda et al. (1997) proposed the empirical capillary-tube relationship (Equation 14). This formulation is commonly referred to as the Tokyo (or Masuda) model. For $N=2$, the model represents a capillary tube wall-coating (pore-coating) gas hydrate morphology, as discussed by Kleinberg et al. (2003). In this work, we estimate gas hydrate saturation from the observed sonic velocity (Figure 6) applying the Biot-Gassmann theory (Ojha and Ghosh, 2021; Mondal et al., 2022) as shown in Figure 7, that demonstrate the estimated hydrate saturation also match well with the available core measurements.

The depth of gas hydrate stability zone (GHSZ) or bottom-simulating reflector (BSR) is approximately 300 mbsf in Area B and 260 mbsf in Area E as estimated from both well log and seismic data (Figure 7). According to the published literatures, the depth of BSR in Area C is 300 mbsf based on seismic data (Collett et al., 2019), despite maximal velocity and resistivity being measured at about 260 mbsf. Figure 8 shows the calculated effective permeability using Equation 14 with taking $N=2$ and calculated absolute permeability using Equation 13., which is also compared with the core measurements that show good match with the calculated estimation.

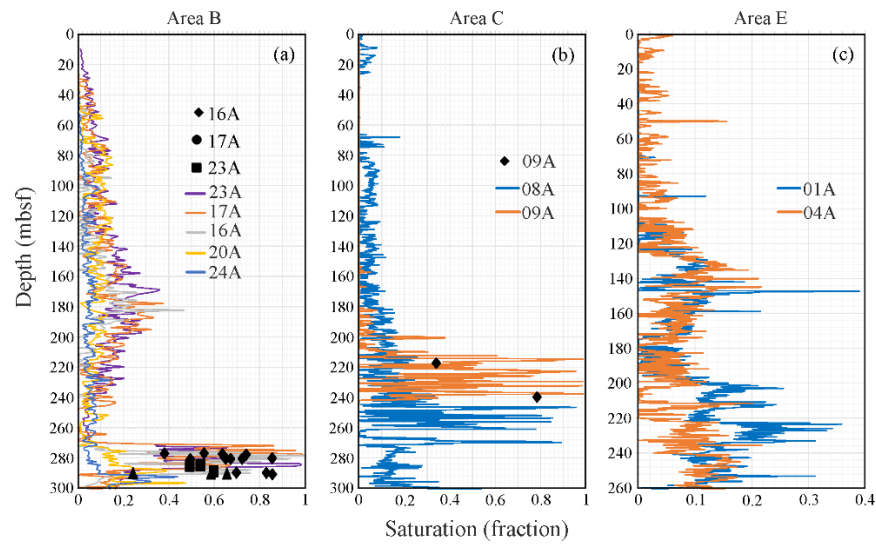


Figure 7. Calculated gas hydrate saturation logs at selected wells in Area B (a), Area C (b), and Area E (c). Solid diamonds, triangles and squares denote the core measurements respectively for wells 16A, 17A, and 23 A.

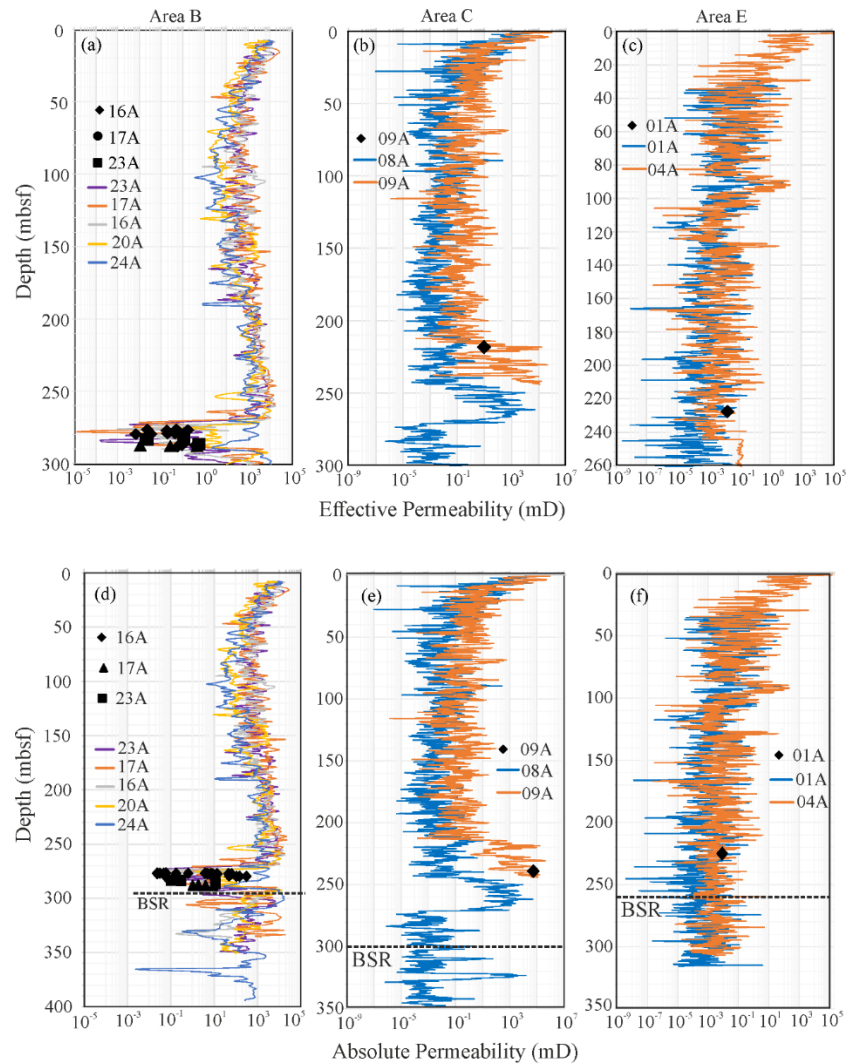


Figure 8. Calculated effective permeability logs at selected wells in areas B, C, and E (a, b, c). Bottom part (d, e, f) shows the absolute permeability of the same for comparison.

Effective permeability and resistivity relation

In this study, we aim to establish the relation between the calculated effective permeability (Figure 8) and the measured resistivity (Figure 6) because of the similarity between the electrical conduction and hydraulic flow, resistivity logs can be used to predict permeability in real-time during drilling. However, we observe that the resistivity and effective

permeability are not well correlated. The correlation improves when regression relation is established between the effective permeability, porosity and resistivity. Figures 9, 10, and 11 show the 3D plots between the resistivity, effective permeability, and porosity for all wells of areas B, C, and E respectively, which provide different power law regression relations between the effective permeability, resistivity and porosity that are displayed in the figures with R-squared values.

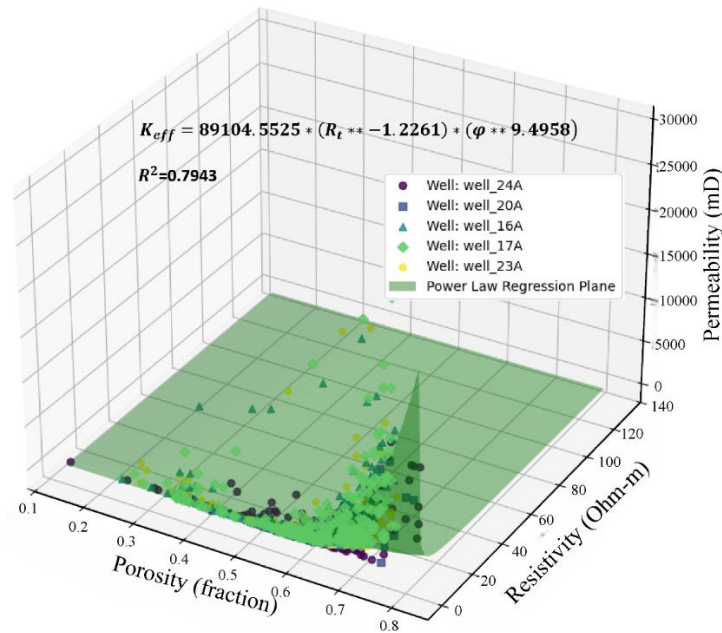


Figure 9. 3D plot between the resistivity, effective permeability, and porosity using the data of the selected wells with power-law-regression relation and R-squared value for Area B.

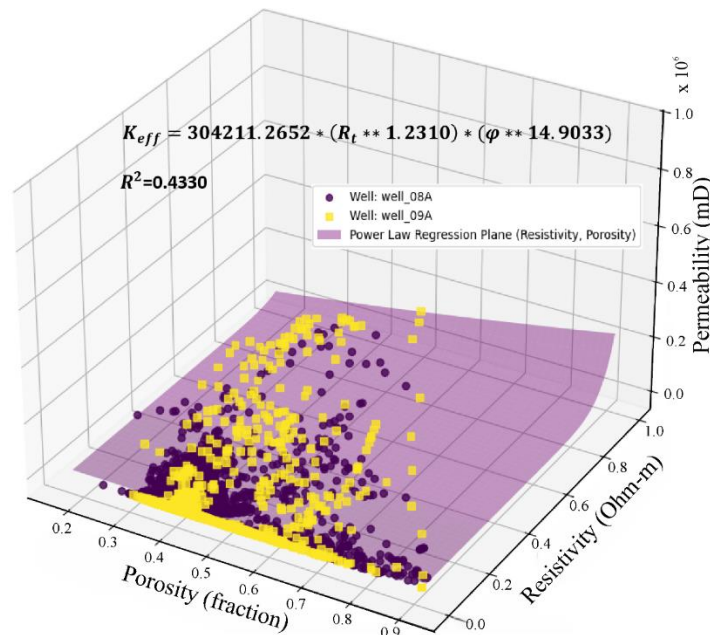


Figure 10. 3D plot between the resistivity, effective permeability, and porosity using the data of the selected wells with power-law-regression relation and R-squared value for Area C.

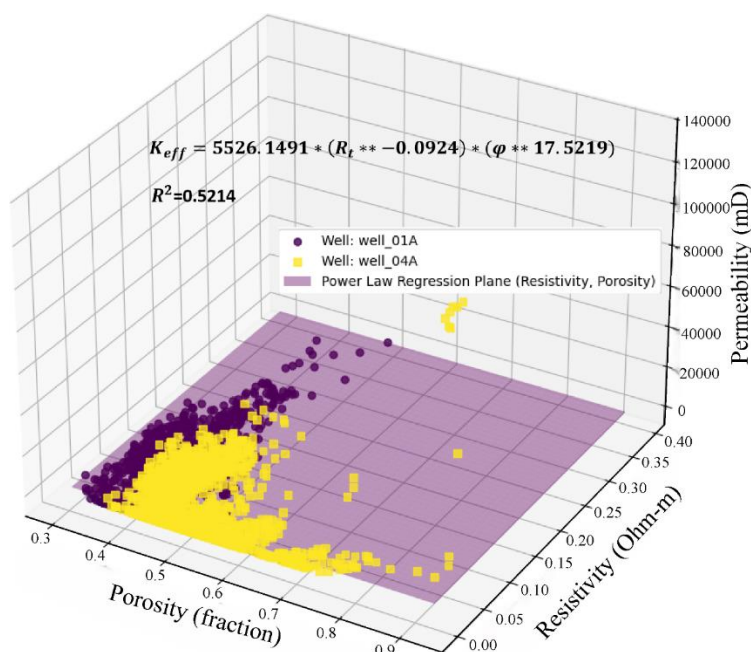


Figure 11. 3D plot between the resistivity, effective permeability, and porosity using the data of the selected wells with power-law-regression relation and R-squared value for Area E.

DISCUSSION

The study aims to derive and compare permeability models of gas hydrate saturated sediments of three identified zones, namely, areas B, C, and E from NGHP-02 in the Krishna-Godavari Basin. We observe that average total porosity is ~60% in Area B, whereas, it is ~40% for Areas C and E. Moreover, average clay volumes in Area B is ~15% and ~60% for Areas C and E. Consequently, the effective porosity decreases by approximately 10% in Area B and by approximately 25% in Areas C and E. In order to establish the best model with the best empirical coefficients to represent these areas, we first examine at the existing empirical relations using the well log data from the areas B, C, and E. This leads to the development of a common empirical relation among the absolute permeability, effective porosity, and S_{wir} for all three areas. The permeability predicted by the derived relation shows good match to the permeability determined from the cores for all areas under investigation.

The effective permeability is more important than the absolute permeability to predict the capacity of gas production in numerical simulators. Therefore, we first estimate the gas hydrate saturation using a well-established rock physics model (Biot-Gassmann theory) from the observed sonic velocity, and then estimate the effective permeability using the given theory as described in the methodology. The results from our methodology show good match with the core measurements for both the predicted gas hydrate saturation and effective permeability. The effective porosity of Area B in the zone of very high gas hydrate saturation (70–80% of the pore space) is

significantly lower than Area C's due to Area C's higher total porosity (~60%) and lower clay volume (<10%) compared to Area B's total porosity (~40%) and clay volume (>20%) in the layers of maximum gas hydrate saturation. This is reflected in the estimated effective permeability of these two regions, where Area B's effective permeability is significantly lower (<10 mD) than Area C's (~100 mD) in the zone of maximum hydrate saturation. However, both total porosity and clay volume in Area E, are much lower in comparison to the other two areas. As a result, Area E has significantly lower effective permeability in spite of having lower gas hydrate saturation than the other two regions.

We analysed the resistivity logs of selected wells in our study as the resistivity logs can be used to predict permeability in real-time while drilling because electrical conduction and hydraulic flow are similar. This is especially helpful in situations where core samples are lost or damaged during retrieval. We find that there is a poor direct correlation between resistivity and effective permeability. Therefore, we generate the 3D plots between the effective permeability, porosity and resistivity to express effective permeability in terms of the total porosity and resistivity of areas B, C, and E. These three parameters are related through different power law regression models for each area with highest R-squared value in Area B (~0.80). Even though porosity and permeability both cannot be directly assessed using the well log tools, porosity is better related to other directly measured well logs (like density log) than permeability. By correlating the effective permeability to resistivity and porosity, it is possible to optimise extraction techniques like depressurisation and anticipate long-term

production more accurately. These correlations aid in determining hydrate growth patterns (e.g., grain-coating vs. pore-filling). For instance, at the same saturation, pore-filling hydrates dramatically decrease permeability more than grain-coating hydrates. This difference can be identified using particular resistivity-permeability exponents. Moreover, the resistivity variations can be utilised to track the dissociation of hydrates in real time during production, assisting in the identification of the displacement of the dissociation front and possible hazards such as sand production.

CONCLUSIONS

This work presents a new permeability estimation model that best captures the gas hydrate-bearing marine sediments of areas B, C, and E in the KG basin. The derived model is validated by available core measurements in all three areas. In Area B, overall absolute and effective permeability are about 1000 mD and 100 mD respectively. In Area C, overall absolute and effective permeability are respectively about 0.1 mD and 0.01 mD. Whereas, in Area E, overall absolute and effective permeability are very less about 0.01 mD and 0.001 mD respectively. We also calculate the gas hydrate saturation and effective permeability to explore the reservoir quality of this basin. According to our study, the effective permeability of a sediment with a gas hydrate saturation of about 80% of its total porosity, is 2% of its absolute permeability. The study reveals that the Area C has the maximum effective permeability in the zones of high gas hydrate saturation. We also provide separate regression relations between the effective permeability, porosity, and resistivity for each area because the resistivity log is a good indicator of hydrate saturation. This is crucial for other locations of the KG basin where it is not possible to measure directly the permeability in the collected core samples. The regression relation in the Area B shows the best correlation of the effective permeability with the porosity and resistivity.

The study's findings indicate that the KG Basin has favourable permeability characteristics for gas production; nonetheless, long-term reservoir integrity is still a fact to consider. Furthermore, although empirical models are frequently basin-specific and rely on the calibration quality of the data, they can aid in the identification of exploration sites. However, machine-learning methods, such as support vector machines and artificial neural networks, offer the benefits of taking spatial variability and uncertainty into account and are better at capturing the intricate nonlinear correlations among reservoir parameters. Our future studies will focus on the relationship between permeability and pore network in gas hydrate reservoirs to gain a better understanding of fluid movement and basin-scale migration pathways.

Acknowledgments

Authors thank Director, CSIR-NGRI for giving permission to publish this work and providing all facilities to carry out research work. We are thankful to the Ministry of Petroleum and Natural Gas (MoPNG) under the Government of India, Directorate General of Hydrocarbons (DGH), New Delhi for providing data. Ministry of Earth Sciences (MoES), Delhi, for providing financial support to carry out the research work on gas hydrate exploration. We sincerely thank the Chief Editor of the Journal of the Indian Geophysical Union (JIGU), and two reviewers for their valuable suggestions, guidance, constructive comments and valuable suggestions, which significantly improved the quality of this manuscript. The Academy of Scientific and Innovative Research (AcSIR), Ghaziabad, Uttar Pradesh, India, is being acknowledged for granting permission to Seema Begum to pursue her Ph.D., and providing necessary academic support.

Author credit statement

Seema Begum: Conceptualization, methodology, data analysis, investigation, writing original draft; Dr. Ranjana Ghosh: Conceptualization, data curation, supervision, validation, writing – review & editing; Dr. Maheswar Ojha: Conceptualization, validation, writing – review & editing, project administration.

Data availability statement

Well log and core data used in this study are available to the corresponding author upon request.

Compliance with ethical standards

The authors affirm that the work presented in the publication was not impacted by any known financial conflicts of interest or personal relationships. They adhere to copyright norms.

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Received on: 15-07-2026; Revised on: 24-07-2026; Accepted on: 03-08-2026

Regime-dependent probabilistic characteristics of annual rainfall in India

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ABSTRACT

Annual rainfall over India exhibits substantial variability and is often treated as approximately Gaussian for interpreting standardized anomalies, probabilistic thresholds, and confidence intervals. However, this assumption has not been systematically evaluated across contrasting climatic regimes. This study analyses the probabilistic structure of annual rainfall, using long-term observational data from three distinct regimes: All-India, Kerala–Mahe and East Rajasthan. A common statistical framework based on Gaussian distribution fitting, standardized Z-score analysis, and confidence intervals is applied uniformly to evaluate distributional characteristics, rainfall anomalies, and uncertainty in the climatological mean. Results show that spatial aggregation at the national scale, produces pronounced Gaussian-like behaviour with suppressed extremes and a stable mean. Kerala–Mahe exhibits a near-Gaussian core with broader but symmetric variability, whereas East Rajasthan displays conditional Gaussianity, characterized by skewness and tail dominance associated with intermittent severe events. Despite marked differences in variability and extremes, the long-term mean remains statistically well constrained across all regions. The study demonstrates that Gaussian rainfall behaviour is scale- and regime-dependent and provides an observational probabilistic framework for interpreting rainfall variability and evaluating climate simulations and data-driven models.

Keywords: Indian rainfall variability; Kerala-Mahe; East Rajasthan; Gaussian-like behaviour; Rainfall extremes; Climatic regimes; Probabilistic analysis

INTRODUCTION

Rainfall over the Indian subcontinent is primarily governed by the summer monsoon and plays a critical role in agriculture, water resources, energy generation, and socio-economic stability. Variations in monsoon rainfall strongly influence food production and water availability, while extreme rainfall events often lead to floods and droughts with significant socio-economic impacts (Parthasarathy et al., 1994; Rajeevan et al., 2006). Consequently, understanding the statistical nature of the rainfall variability and extremes, remains a central problem in climatological research.

Indian monsoon rainfall has been studied extensively using long-term observational records, numerical climate models, and more recently, data-driven and machine-learning approaches (Webster et al., 1998; Rajeevan, 2001; Goswami et al., 2006). Much of this work focuses on prediction, attribution, or trend detection, often with emphasis on large-scale climate drivers such as ENSO and the Indian Ocean Dipole. While these approaches have advanced understanding of monsoon dynamics, they frequently rely, either explicitly or implicitly, on assumptions regarding the underlying statistical structure of rainfall, particularly at annual time scales. At seasonal and annual scales, rainfall anomalies are commonly treated as approximately Gaussian. This assumption underpins the widespread use of standardized anomalies, confidence intervals, and probabilistic thresholds for identifying extreme events (von Storch and Zwiers, 1999; Wilks, 2011). Studies based on All-India rainfall have shown that spatial aggregation suppresses local variability and stabilizes the long-term mean, leading to near-Gaussian behaviour in annual totals

(Parthasarathy et al., 1990; Rajeevan et al., 2007). As a result, Gaussian statistics are often adopted as a convenient and seemingly reasonable approximation in climatological analyses.

At regional scales, however, rainfall behaviour can deviate substantially from this idealized picture. Several studies have reported skewness, asymmetry and heavy-tailed behaviour in rainfall distributions, particularly in regions influenced by orography or characterized by intermittent rainfall (Guhathakurta and Rajeevan, 2008; Ghosh et al., 2012). More recent work continues to emphasize the complexity of rainfall extremes and their regional diversity across India, underscoring the need for careful statistical characterization based on observations rather than assumptions alone (Samantray and Gouda, 2024; Joseph et al., 2024; Veeram et al., 2026).

Despite this extensive studies, the validity of Gaussian-like assumptions across contrasting climatic regimes is often taken for granted rather than explicitly examined using a consistent statistical framework. In particular, three related but conceptually distinct aspects are frequently merged: (i) the distributional structure of rainfall totals, (ii) the character of standardized anomalies with respect to mean and temporal extremes or spikes, and (iii) the uncertainty associated with the estimated climatological mean (Wilks, 2011). High interannual variability is sometimes interpreted as instability of the mean climate state, even when long observational records provide a statistically well-constrained estimate of the mean.

From a statistical perspective, annual rainfall may be viewed as a stochastic variable resulting from the integrated effects of numerous atmospheric and oceanic processes. Under sufficient

spatial or temporal aggregation, approximate Gaussian behaviour may emerge in the core of the distribution through averaging, while departures from Gaussianity may persist in the form of skewness or tail dominance, particularly in regions subject to strong intermittency or localized extremes (von Storch and Zwiers, 1999; Ankush et al., 2024). Explicitly demonstrating where such Gaussian-like behaviour holds, and where it breaks down, remains an important but relatively underexplored problem in observational climatology.

In this study, the probabilistic structure of annual rainfall is examined using long-term observational data and a unified statistical framework. Gaussian distribution fitting is employed to assess the extent to which annual rainfall exhibits Gaussian-like behaviour in the core and tails of the distribution. Standardized Z-score analysis is used to characterize the frequency, symmetry, and magnitude of rainfall anomalies, enabling an objective identification of extreme years. Confidence intervals for the long-term mean rainfall are computed to distinguish natural interannual variability from uncertainty in the estimated mean. The analysis is applied uniformly to three contrasting climatic regimes: All-India rainfall, representing strong spatial aggregation; Kerala–Mahe, a humid, ocean-facing region influenced by orography; and East Rajasthan, a semi-arid region characterized by intermittent rainfall and occasional severe extremes, as defined by the India Meteorological Department. By applying identical statistical tools across these regions, the study explicitly demonstrates how probabilistic rainfall characteristics emerge, weaken, or break down as a function of aggregation and climatic regime. The contribution of this work lies not in introducing new statistical methods, but in demonstrating—rather than assuming—the regime dependence of Gaussian-like rainfall behaviour, and in clearly separating variability, extremes, and uncertainty within a single, transparent framework. Recent work has highlighted the importance of evaluating rainfall prediction models in terms of their ability to preserve fundamental statistical characteristics of observed rainfall time series (Goyal et al., 2025). This study complements this perspective by establishing a regime-aware probabilistic baseline for interpreting rainfall variability and extremes, providing a reference for future numerical and machine-learning-based studies.

DATA AND METHODOLOGY

Rainfall data

The analysis is based on long-term rainfall records obtained from the India Meteorological Department (IMD) (<https://mausam.imd.gov.in/>). Annual rainfall time series for the period 1901–2023 were constructed for three spatial domains: (i) All-India, (ii) Kerala–Mahe, and (iii) East

Rajasthan. The Kerala–Mahe series corresponds to the official IMD meteorological subdivision comprising the state of Kerala along with the Mahe enclave of Puducherry, which is geographically embedded within Kerala and shares the same monsoon rainfall regime. Monthly rainfall values, expressed in millimetres (mm), were aggregated to annual totals by summing rainfall over the twelve calendar months for each year. Only complete calendar years were retained in the analysis to ensure consistency across regions. Annual rainfall is treated as a realization of a stochastic process, where each year’s rainfall total represents an outcome influenced by numerous interacting atmospheric and oceanic factors. For each region, the annual rainfall sequence is therefore regarded as a sample drawn from an underlying probability distribution.

Derivation of regional rainfall from IMD observations

Rainfall observations used in this study are based on surface rain-gauge measurements collected and quality-controlled by the India Meteorological Department. IMD maintains a dense network of rain-gauge stations distributed across India, providing long-term records of daily rainfall. For each station, daily rainfall amounts are recorded using standard rain-gauge instruments and subjected to routine quality checks to identify missing data and observational inconsistencies. Monthly rainfall totals are obtained by summing daily rainfall values within each calendar month. Annual rainfall at a station is then computed as the sum of rainfall over the twelve calendar months.

Regional rainfall series are derived by spatially aggregating station-level annual rainfall over predefined meteorological subdivisions. The All-India rainfall series represents an area-weighted aggregation of rainfall over all subdivisions, whereas the Kerala–Mahe and East Rajasthan series, correspond to their respective IMD-defined subdivisions. This spatial aggregation implicitly smooths local-scale variability while preserving regional-scale rainfall characteristics. The resulting annual rainfall series, therefore, represents a climatologically consistent estimate of regional mean rainfall, suitable for long-term statistical analysis of distributional structure, anomalies, and variability. Daily rainfall observations from IMD’s network of approximately 7000 stations are aggregated to estimate daily rainfall for each of the 36 meteorological subdivisions. R_s can be written as,

$$R_s = \frac{\sum_i R_i}{N_s} \quad (1)$$

where R_i is the rainfall of one day for the i^{th} station of the s^{th} subdivision. N_s is the number of reporting stations within subdivision s . R_s is the rainfall of the subdivision s for a single

day. The monthly rainfall for the subdivision s and month m is obtained as:

$$R_{m,s} = \sum_{\text{No of days}} R_s \quad (2)$$

where $R_{m,s}$, is the monthly rainfall of subdivision s for month m .

The monthly All India Rainfall (AIR) is calculated from this monthly subdivisional data by weighted average over area of subdivisions:

$$R_{m,AIR} = \frac{1}{(\sum_s A_s)} \sum_s (A_s R_{m,s}) \quad (3)$$

where $R_{m,AIR}$ is the AIR for the month m , and A_s is the area of the s^{th} subdivision. The monthly regional rainfall is calculated in similar manner.

Rainfall as a stochastic variable

In this study, the annual rainfall values are treated as samples of a random variable R . From eqn. (3), the annual rainfall is calculated by adding the precipitation of 12 months for that year.

$$R_t = \sum_m R_{m,AIR} \quad (4)$$

Here, R_t denotes the annual All-India rainfall. Since the data set is of 123 years, we get random observations set as:

$$R = \{R_1, R_2, \dots, R_{123}\} \quad (5)$$

It is this random variable R that we tend to model. A histogram of these samples is first constructed to visualize the empirical probability distribution. The sample mean and standard deviation, are considered as estimates of the expected value and standard deviation of the random variable R , respectively.

The sample mean μ and sample standard deviation σ are computed as

$$\mu = \frac{1}{123} \sum_{i=1:123} R_i \quad (6)$$

$$\sigma = \left[\frac{1}{123} \sum_{i=1:123} (R_i - \mu)^2 \right]^{1/2} \quad (7)$$

These quantities provide estimates of the central tendency and dispersion of the rainfall distribution for each region.

The annual All-India rainfall time series (1901–2023) yields an empirical sample mean (μ) of 1179 mm and an empirical sample standard deviation (σ) of 107.89 i.e. about 11%. The corresponding sample mean and standard deviation are

approximately 668.6 mm and 165.14 mm for East Rajasthan, and 2918.1 mm and 430.8 mm for Kerala–Mahe, respectively. Under the assumption of stationarity and Gaussian behaviour of annual rainfall, this value also represents the expected long-term mean rainfall $E[R]$ of the underlying climatic process (Wilks, 2011).

Gaussian distribution fitting

To examine the distributional structure of annual rainfall, a Gaussian probability density function was fitted to each regional dataset using the estimated mean μ and standard deviation σ . The fitted distribution is given by

$$f(R) = \frac{1}{\sigma\sqrt{2\pi}} \exp\left(-\frac{(R-\mu)^2}{2\sigma^2}\right) \quad (8)$$

Normalized histograms of annual rainfall were constructed and compared with the fitted Gaussian curves to assess the extent to which the empirical distributions exhibit Gaussian-like behaviour. Emphasis is placed on the correspondence between the fitted curve and the core of the distribution, rather than on strict statistical acceptance or rejection. In this study, normalization refers to scaling the histogram such that the total area equals unity, enabling comparison with probability density functions. In contrast, standardization refers to the transformation of rainfall values into Z-scores using the sample mean and standard deviation.

Standardized anomaly (Z-score) analysis

Standardized rainfall anomalies were computed using a Z-score transformation to express deviations from the long-term mean in units of standard deviation:

$$Z_t = \frac{R_t - \mu}{\sigma} \quad (9)$$

where Z_t is the standardized anomaly in year t . This transformation allows direct comparison of anomaly magnitude across regions with different rainfall climatologies. In this study, Z-scores are used to characterize the frequency, symmetry, and magnitude of rainfall anomalies. Years with $|Z_t| \geq 2$ are considered statistically extreme, following conventional climatological practice.

Confidence interval for the climatological mean

To quantify uncertainty in the estimated long-term mean rainfall, a 95% confidence interval was computed for each regional mean under the assumption of independent annual observations:

$$\mu \pm 1.96 \frac{\sigma}{\sqrt{N}} \quad (10)$$

where μ and σ denote the sample mean and standard deviation, respectively, and N is the number of observations. The factor 1.96 corresponds to the 95% confidence level under the

assumption of approximate normality. This expression defines the 95% confidence interval for the estimated mean rainfall, representing the uncertainty associated with estimating the true long-term mean from a finite sample. The interval is used to distinguish interannual variability from uncertainty in the mean, rather than to test for temporal change.

Interpretation of Gaussian-like behaviour

Gaussianity is interpreted here in a pragmatic climatological sense rather than through formal hypothesis testing. A rainfall distribution is considered *Gaussian-like* if it exhibits approximate symmetry in the core of the distribution, balanced standardized anomalies about the mean, and tail behaviour that does not dominate the overall variance. Departures from Gaussianity are therefore assessed qualitatively in terms of skewness, anomaly asymmetry, and tail dominance

RESULTS

Distributional characteristics of annual rainfall

Figure 1 illustrates the empirical annual rainfall distributions and corresponding Gaussian probability density functions for All-India (a), East Rajasthan (b), and Kerala–Mahe (c) over 1901–2023. At the macro scale, All-India rainfall (Figure 1a) exhibits a highly symmetric, well-defined bell-shaped structure closely aligning with the Gaussian fit, with deviations constrained to the extreme tails. Conversely, regional regimes show distinct departures from normality. East Rajasthan (Figure 1b) displays weaker Gaussian correspondence, marked by noticeable asymmetry and heavier right-hand tails that reflect high-magnitude, intermittent events. Kerala–Mahe (Figure 1c), shows a broad but largely symmetric core centered around its mean; the Gaussian curve captures the central spread well despite tail departures induced by higher regional variance. Cross-regional differences in core symmetry, tail behaviour, and overall conformity to Gaussian expectations are summarized in Table 1. Together, Figure 1 and Table 1 demonstrate how Gaussian-like behaviour systematically weakens away from large-scale spatial aggregations.

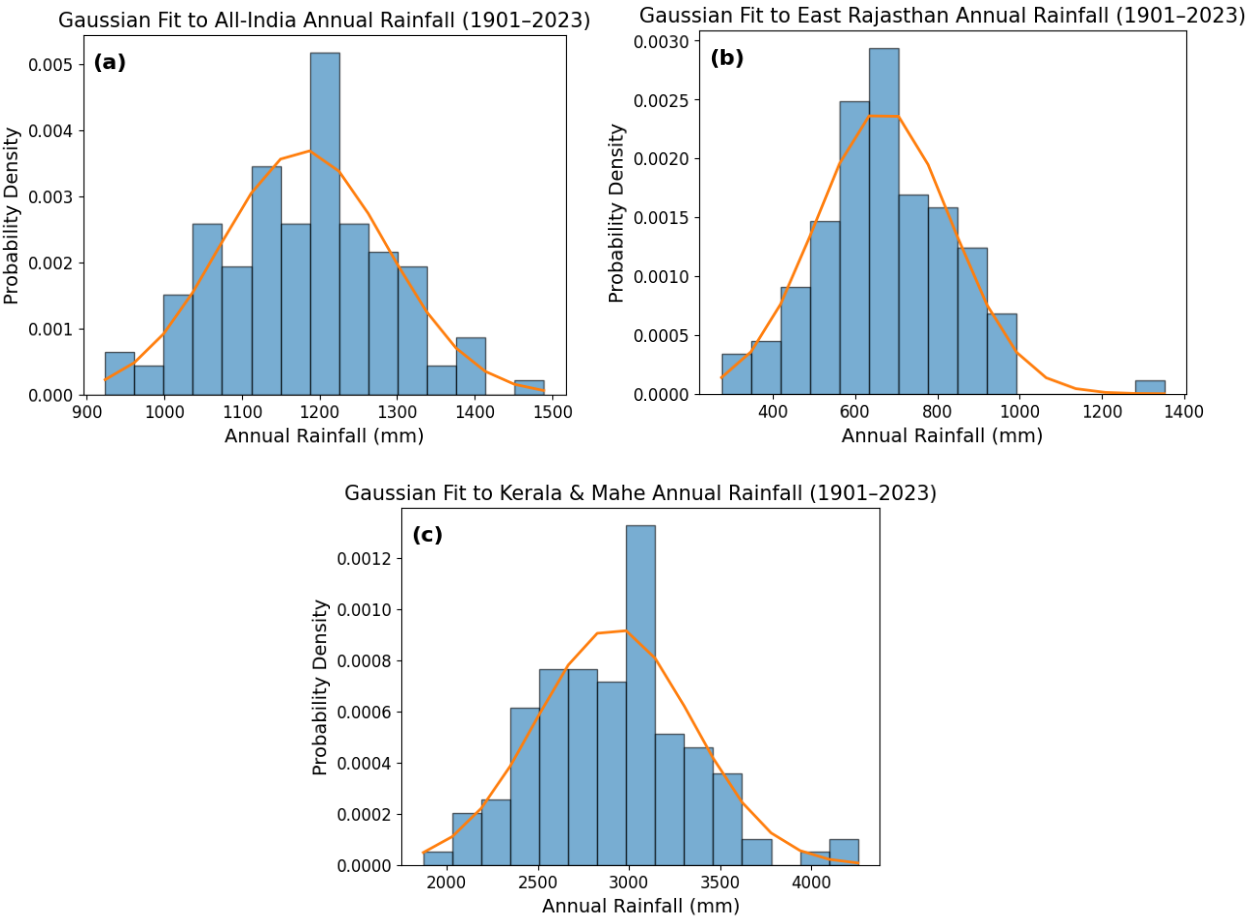


Figure 1. Gaussian fit to the distribution of annual rainfall for (a) All-India, (b) East Rajasthan, and (c) Kerala–Mahe during 1901–2023. The histograms represent normalized rainfall frequency, while the solid curves denote the fitted Gaussian probability density functions based on the sample mean and standard deviation. The comparison highlights the emergence of near-Gaussian behaviour under spatial aggregation and its progressive weakening across distinct climatic regimes.

Standardized rainfall anomalies and extreme-event characteristics

Figure 2 presents the time series of standardized annual rainfall anomalies (Z-scores) for (a) All-India, (b) East Rajasthan and (c) Kerala–Mahe over 1901–2023. Horizontal reference lines at $\pm 1\sigma$ and $\pm 2\sigma$, indicate moderate and extreme departures from the climatological mean, respectively. Visually apparent downward tendencies in parts of the time series are statistically insignificant, remaining within natural interannual variability bounds. At the national scale (Figure 2a), most years fall within $\pm\sigma$, with only a small number of years exceeding $\pm 2\sigma$ threshold. Positive and negative anomalies are broadly balanced. East Rajasthan (Figure 2b) displays an asymmetric anomaly structure; while most of the years remain within $\pm\sigma$, it features severe, isolated positive extremes- including a singular event exceeding 4σ - with comparatively muted negative deviations. In contrast, Kerala–Mahe (Figure 2c) experiences more frequent, temporally balanced excursions beyond $\pm 2\sigma$ on both sides, though anomalies generally remain approximately 3σ . A comparative summary of the frequency, symmetry, and magnitude of standardized rainfall anomalies across the three regions is also presented in Table 2. Combined, Figure 2 and Table 2 demonstrate clear regime-dependent characteristics in the distribution and symmetry of rainfall extremes.

Interannual variability and uncertainty in the climatological mean

Figure 3 presents the annual rainfall time series (1901–2023) for (a) All-India, (b) East Rajasthan, and (c) Kerala–Mahe with their long-term means (μ) and shaded 95% confidence intervals ($\mu \pm 1.96\sigma/\sqrt{N}$). This visualization explicitly decouples interannual variability (year-to-year fluctuations) from climatological mean uncertainty (the statistical precision of the estimated average). At the macro scale, the All-India series (Figure 3a) ($\mu \approx 1179\text{mm}$, $\sigma \approx 107\text{mm}$, $N = 123$) yields a narrow confidence interval of 1160–1198 mm, indicating that the mean is tightly constrained despite volatile annual swings. Regionally, despite localized extreme anomalies, high statistical precision is preserved. East Rajasthan (Figure 3b) exhibits pronounced relative variability, yet its confidence interval remains remarkably narrow (639–698 mm), demonstrating that severe, isolated events do not destabilize the long-term baseline. Similarly, Kerala–Mahe (Figure 3c) reflects a high-amplitude, monsoon-dominated regime, but maintains a well-constrained average interval of 2842–2994 mm. Table 3 summarizes these comparative metrics. Collectively, Figure 3 and Table 3 show that progressing from national to regional scales, the confidence interval expands slightly as a direct function of increased standard deviation (σ); however, across all regimes, the intervals remain tight relative to their means, confirming highly stable climatological baselines.

Table 1. Summary of distributional characteristics inferred from Gaussian fitting of annual rainfall for All-India, East Rajasthan, and Kerala–Mahe. The table qualitatively assesses core symmetry, tail behaviour, and overall Gaussian-like characteristics for each climatic regime based on Figure 1.

<i>Region</i>	<i>Core Gaussianity</i>	<i>Symmetry</i>	<i>Tail behaviour</i>	<i>Overall</i>
All-India	Strong	Strong	Mild deviation	Most Gaussian
Kerala–Mahe	Moderate–Strong	Strong	Broad but symmetric	Moderately Gaussian
East Rajasthan	Moderate	Weak	Skewed, tail-dominated	Weakly Gaussian

Table 2. Summary of standardized anomaly behaviour for All-India, East Rajasthan, and Kerala–Mahe, including the relative frequency, symmetry, and magnitude of extreme ($|Z_t| \geq 2$) rainfall events inferred from Figure 2.

<i>Region</i>	<i>Frequency of $\pm 2\sigma$ events</i>	<i>Maximum</i>	<i>Symmetry</i>	<i>Character of extremes</i>
All-India	Very low	$< 3\sigma$	Strong	Suppressed
Kerala–Mahe	Moderate	$\approx 3\sigma$	Strong	Frequent–moderate
East Rajasthan	Low	$> 4\sigma$	Weak	Rare–severe

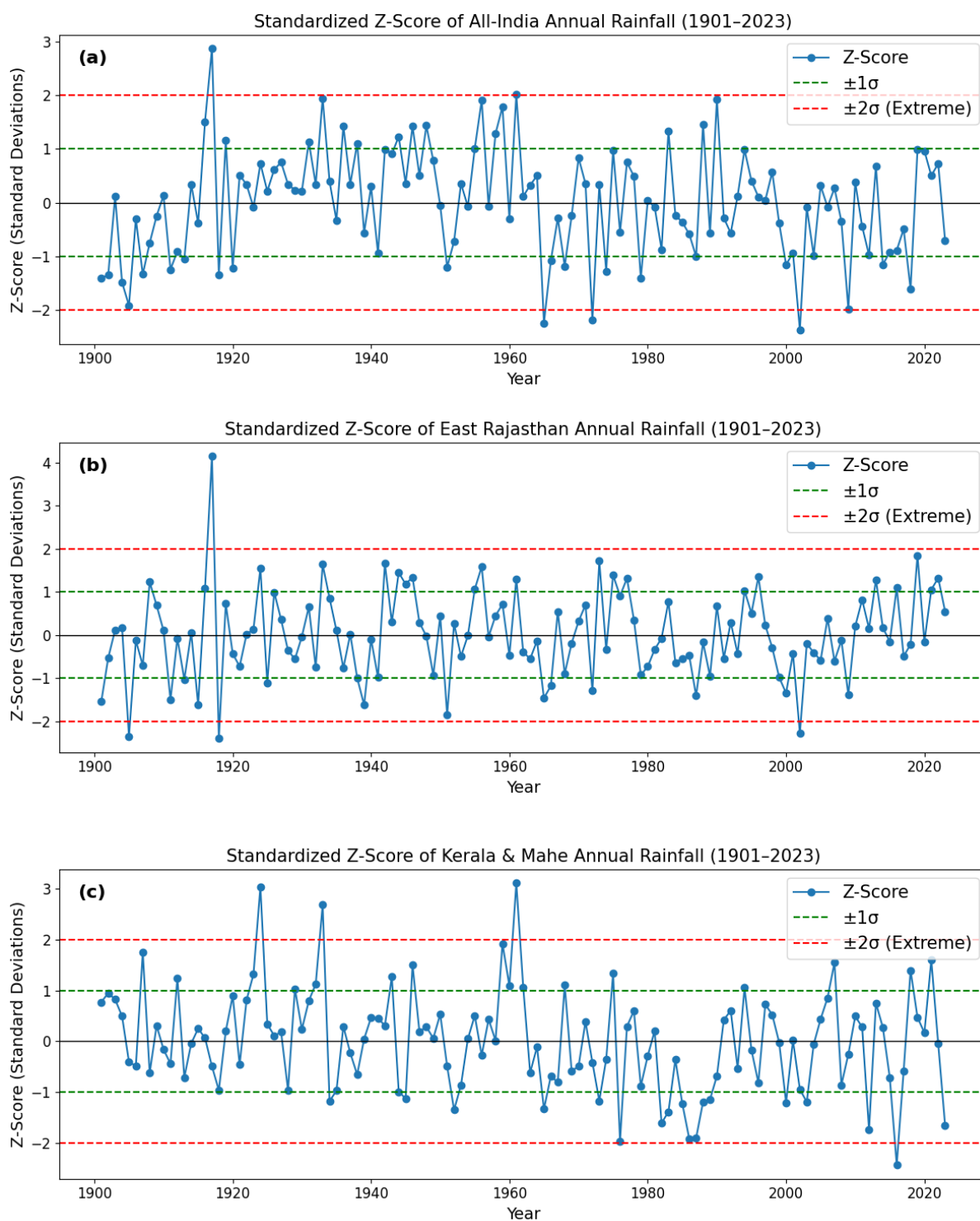


Figure 2. Time series of standardized annual rainfall anomalies (Z-scores) for (a) All-India, (b) East Rajasthan, and (c) Kerala–Mahe during 1901–2023. Horizontal dashed lines indicate $\pm 1\sigma$ and $\pm 2\sigma$ thresholds. The figure illustrates regime-dependent differences in the frequency, symmetry, and magnitude of extreme rainfall anomalies.

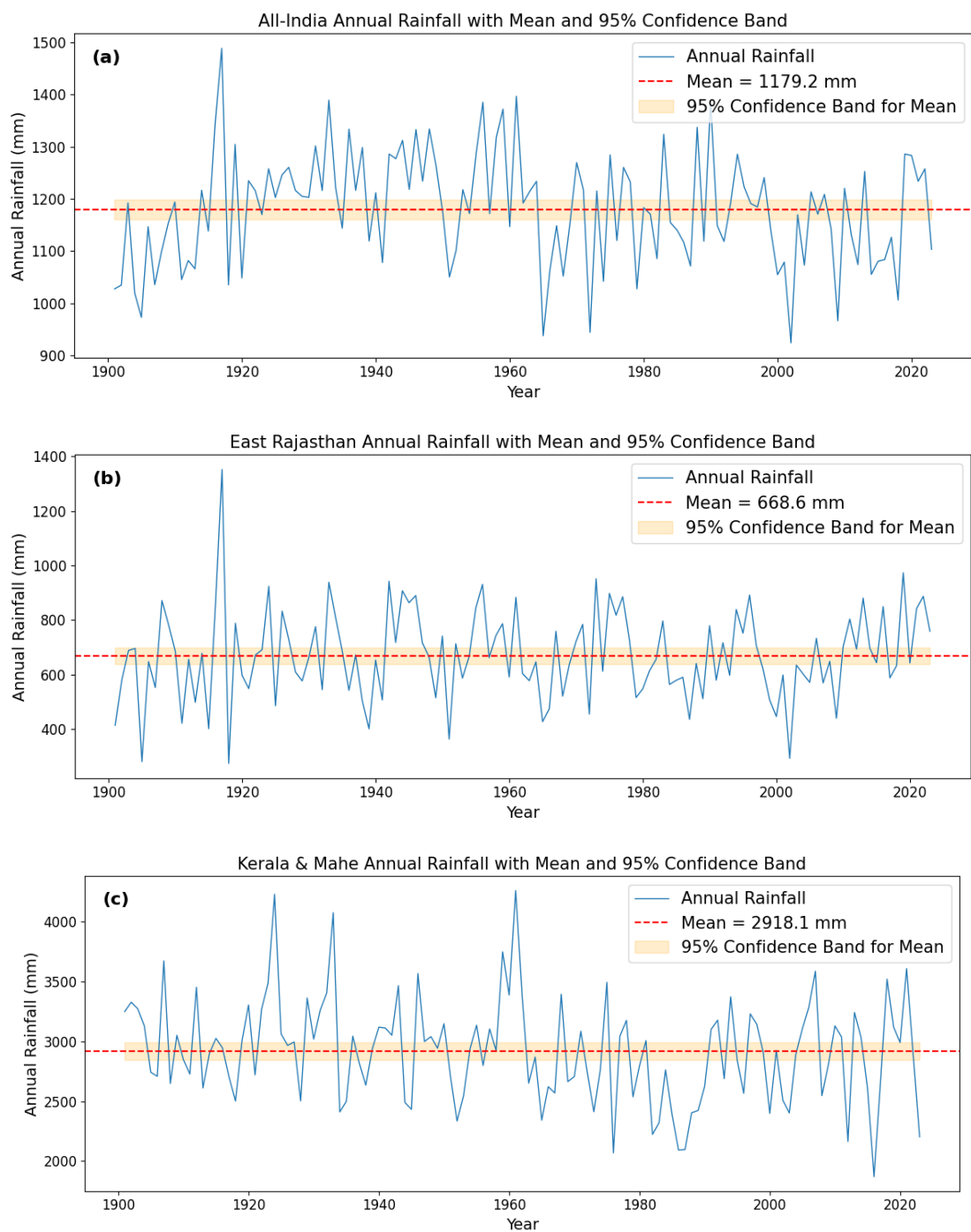


Figure 3. Annual rainfall time series for (a) All-India, (b) East Rajasthan, and (c) Kerala–Mahe during 1901–2023, shown together with the long-term mean (dashed line) and its 95% confidence interval (shaded band). The figure demonstrates that, despite substantial interannual variability and extremes, the long-term mean rainfall remains statistically well constrained across contrasting climatic regimes.

Table 3. Comparison of interannual rainfall variability and uncertainty in the estimated long-term mean for All-India, East Rajasthan, and Kerala–Mahe, as inferred from the 95% confidence intervals shown in Figure 3.

Region	Variability	Extremes	CI width	Mean stability
All-India	Moderate	Rare	Narrow	High
Kerala–Mahe	High	Moderate	Narrow	High
East Rajasthan	High	Severe (rare)	Narrow	High

DISCUSSION

This study examined the probabilistic structure of annual rainfall across three contrasting climatic regimes, All-India, Kerala–Mahe, and East Rajasthan, using a unified framework comprising Gaussian distributional analysis, standardized anomalies, and confidence intervals for the climatological mean. Together, the results provide a coherent picture of how statistical structure emerges under aggregation and degrades under intermittency, while remaining sufficiently stable to support objective definitions of anomalies and extremes.

Emergence, scale, and mechanics of regimes

This study demonstrates that the probabilistic structure of Indian annual rainfall is neither binary nor universal, but emerges or degrades as a function of spatial aggregation and regional climate mechanics. At the macro scale, extensive national aggregation suppresses localized variance, generating a highly symmetric, Gaussian-like distribution. Regionally, the underlying physical mechanisms dictate distinct structural departures. Kerala–Mahe's humid, orographically driven regime maintains a near-Gaussian core despite high variance, indicating that elevated variability alone does not disrupt normality. Conversely, East Rajasthan exhibits only conditional Gaussianity; its central core conforms to a normal approximation, but intermittency introduces pronounced right-hand skewness and tail dominance.

Frequency versus magnitude of extremes

The regional divergence clarifies the distinction between extreme frequency and extreme magnitude. In Kerala–Mahe, extremes ($|Z_t| \geq 2$) are frequent, moderate, and symmetric, reflecting continuous high variability. In East Rajasthan, extremes are rare but severe, featuring major singular excursions ($|Z_t| \geq 4$). Thus, capturing a regime's extreme-event character requires looking beyond a simple count of anomalous years to evaluate the directional asymmetry of the tails. All-India rainfall displays strongly suppressed extremes, consistent with spatial averaging.

Variability versus uncertainty in the mean

A critical finding is that high interannual variability does not imply uncertainty in the climatological mean. Across all three regimes, long observational records yield narrow confidence bands, confirming that the identified Z -score anomalies represent genuine climatic deviations rather than statistical artifacts of baseline instability. This clear separation between natural year-to-year volatility and mean precision provides a robust foundation for defining true climate extremes.

Implications and broader relevance

Consequently, evaluating rainfall distributions must move past binary hypothesis testing toward a pragmatic assessment of core symmetry, anomaly balance, and tail behaviour. While this study focuses on observation, this regime-aware baseline provides a vital benchmark for validating numerical and machine learning prediction frameworks. Rather than relying exclusively on standard error-based metrics, future models must be rigorously evaluated on their capacity to preserve these observed distributional shapes and scale-dependent extreme structures to ensure true physical realism. Furthermore, this probabilistic framework can be extended to evaluate the stochastic behaviour of other vital monsoon characteristics, such as seasonal onset dates.

CONCLUSIONS

1. Our study shows strong spatial aggregation, as in the All-India rainfall series, leads to pronounced Gaussian-like behaviour, with suppressed extremes and a stable mean. Kerala–Mahe, representing a humid and orographically influenced regime, exhibits a near-Gaussian core with broader but largely symmetric variability, resulting in more frequent moderate extremes. In contrast, East Rajasthan displays only conditional Gaussianity, characterized by skewness and tail dominance arising from intermittent but occasionally severe rainfall events. These findings confirm that Gaussianity in rainfall is neither universal nor binary, but emerges or degrades in a regime-dependent manner.
2. Standardized anomaly analysis further revealed that extreme rainfall behaviour differs fundamentally across regions, not only in frequency but also in magnitude. While Kerala–Mahe is characterized by more frequent moderate anomalies, East Rajasthan experiences rarer but more severe deviations. Despite these differences, confidence-interval analysis showed that the long-term mean rainfall remains statistically well constrained across all regions, underscoring the distinction between natural interannual variability and uncertainty in the estimated mean.
3. Beyond their climatological significance, these results have direct implications for the development and evaluation of machine-learning-based rainfall prediction models. Data-driven models are often assessed primarily using error-based performance metrics, without explicit consideration of whether they reproduce the observed probabilistic structure of rainfall.
4. By clarifying what constitutes *normal variability* and *statistically extreme behaviour* across different climatic

regimes, the present framework can guide the design, training, and validation of machine-learning models toward improved physical realism. In particular, models that are able to reproduce not only mean rainfall but also the observed regime-dependent anomaly structure, are more likely to yield robust and interpretable predictions.

Author credit statement

Pooja Mishra: Investigation, coding, analysis and initial draft of the manuscript; K. C. Tripathi: Conceptualization, methodology, supervision, and critical revision of the manuscript.

Data availability

The rainfall data used in this study are publicly available from the India Meteorological Department (IMD) website.

Compliance with ethical standards

The authors declare that they have no conflict of interest and adhere to copyright norms.

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Received on: 27-1-2026; Revised on: 12-06-2026; Accepted on: 26-07-2026

Recent surge in seismic activity and neo-tectonic reactivation of faults in Deccan Volcanic Province (India) and adjoining regions

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ABSTRACT

The recent increase in intraplate seismicity across the Deccan Volcanic Province and adjoining parts of central and western Peninsular India, has raised concerns regarding the growing seismic hazard and public safety. The occurrence of the 2024 Mulugu earthquake in Telangana, followed by a recent earthquake sequence in Nanded-Parbhani-Hingoli regions of Maharashtra, together with the seismic activity across Koyna-Nashik-Palghar region, suggest that the intraplate deformation is not spatially restricted to local seismic zones, but may reflect a broader regional stress regime. To investigate the temporal evolution of intra-plate seismicity, we compiled and compared earthquake events of magnitude M: 1-8 and 4-8 separately for the period 1950-2026, using successive 10-year time window. The present analysis revealed a pronounced increase in the frequency and spatial distribution of earthquakes through time, evident in both the complete M: 1-8 catalogue and the relatively higher-magnitude M: 4-8 events. The progressive increase in seismicity over the past 5-7 decades is particularly noteworthy, suggesting that the observed recent activity represents more than a short-term fluctuation. This temporal pattern, together with the spatial clustering of earthquakes along inherited rift and fault-zones, indicates a sustained increase in intraplate deformation and reactivation of pre-existing crustal weakened zones across the Deccan volcanic terrain and adjoining regions. Although, improved seismic-network coverage and detection capabilities may contribute to the apparent increase, the persistence of the trend in the M: 4-8 population, suggest a considerable component of increasing tectonic activity. The recent seismicity may be interpreted as the combined effect of far-field Himalayan compressional stresses, intra-plate stress redistribution, and reactivation of inherited faults associated with the Precambrian and Proterozoic rift systems. We further propose that the asthenospheric thermal upwelling beneath the Deccan region, manifested by a number of uplifted domains, may contribute to the crustal weakening, thereby modifying the regional stress field, resulting into the reactivation of pre-existing fault zones.

INTRODUCTION

Peninsular India constitutes one of the oldest and arguably the most tectonically stable continental region of the Indian plate having witnessed relatively lesser seismic activity. Nevertheless, the region has experienced several moderate to large intraplate earthquakes throughout historical and instrumental times, demonstrating that the apparently stable continental lithosphere remains capable of accumulating and releasing significant tectonic stresses (Gupta et al. 2001; Vasanthi, 2022). Notable examples include, the 1819 Kutch earthquake (M ~7.7), the 1927 Son Valley earthquake (M ~6.4), the 1967 Koyna earthquake (M ~6.3), the 1969 Bhadrachalam earthquake (M ~5.7), the 1993 Killari earthquake (M ~6.2), the 1997 Jabalpur earthquake (M ~5.8), and the 2001 Bhuj earthquake (Mw 7.7). More recently, moderate earthquakes in central and southern Peninsular India, including the 2024 Mulugu earthquake (Vasanthi, 2025), and the 2026 Hingoli seismic sequence, indicate that intraplate deformation remains active within this supposedly stable lithospheric domain.

Usually, major intraplate earthquakes, are known to have intimate relationships with rift valleys, like New Madrid (USA), Baikal (Siberia), East African region, Shanxi rift zone in China and Koyna, Kurduwadi, Narmada - Son, Kutch and Godavari in India to name a few (Liu and Zoback, 1997; Mandal and Pandey, 2010; Catchings et al., 2015; Vasanthi and Satish Kumar, 2016; Mukherjee et al., 2017; Oliva et al., 2019; Xu et al., 2020; Vasanthi, 2022, 2025). The recent Hingoli activity is particularly significant when considered in a broader context of seismicity across central and southern peninsular India, which include the earthquake sequences documented around Koyna-Nashik-Palghar, Killari, Nanded and Parbhani regions. These intraplate earthquakes are also seen to occur

within or close to a network of ancient crustal weak zones, that include rift systems and associated basement faults, possibly indicating that the present-day seismicity, may be controlled at least in part, by the mechanical reactivation of ancient crustal discontinuities, caused by neo-tectonic activities (Vasanthi, 2025).

A temporal analysis of the seismic catalogue provides an additional perspective on the intra-plate seismicity. In the present study, earthquakes of M: 1-8 and M: 4-8 are analysed separately for the period 1950-2026, using successive 10-year window for the Deccan volcanic and adjoining regions. The temporal distribution indicates a pronounced increase in the number and spatial extent of recorded earthquakes during the last five to seven decades. Although part of this increase may be attributed to improvements in seismic-network coverage and detection capabilities, the persistence of moderate-magnitude events, provides evidence for a progressive change in the seismic behaviour of the region. The increasing concentration of earthquakes around established structural corridors, further suggests that the observed seismicity is not entirely random, but may be linked to the progressive reactivation of inherited crustal weak zones. We further explore whether this seismicity can be explained by the combined influence of far-field Himalayan compression, and possible asthenospheric thermal upwelling beneath the Deccan region.

TEMPORAL VARIATION IN SEISMICITY IN MAHARASHTRA AND ADJOINING REGIONS

To investigate the temporal evolution of seismicity across the Deccan volcanic and adjoining regions, earthquake events

taken from National Centre for Seismology (NCS) Data Portal, were analysed in successive 10-year time window, beginning with 1950-1960 (Figure 1) and continuing through 2020, followed by a final window of 2020-2026 to capture the most recent seismic activity. The earthquakes were examined separately in two magnitude ranges, M : 4-8 and M : 1-8 (Figures 1-8), in order to distinguish the temporal behaviour of the complete recorded seismicity from that of minor, moderate and large events. For each window, the spatial distribution of earthquakes was mapped to evaluate changes in both frequency and geographic extent of seismic activity. The earliest time window, 1950-1960, represents a period of very limited documented seismicity. Only three earthquakes of $M \geq 4$ were identified during this interval (Figure 1), and the spatial distribution of events in the M : 4-8 and M : 1-8 categories are essentially identical. The similarity may possibly reflect the limitations of earthquake detection during the early period, when seismic observations were primarily based on analogue recording systems, and the detection of low-magnitude events was considerably more restricted than in the modern instrumental era. Consequently, the apparent absence of

smaller earthquakes during this period (Figure 1b) should not necessarily be interpreted as an actual absence of micro-seismicity. Rather, it represents the detection threshold and observational limitations of the available seismic network.

During the subsequent decade, 1960-1970 (Figure 2), an increase in the number of earthquakes of $M \geq 4$ becomes evident, compared with the 1950-1960 interval, which include 1967 Koyna earthquake. This increase continues through the following decades (Figures 3 and 4), although the magnitude and spatial distribution of the events vary considerably between individual time windows. Importantly, up to approximately 1990, the spatial patterns represented by the M : 4-8 and M : 1-8 earthquake maps, remain the same. This similarity indicates that the available earthquake catalogue was dominated by events above the magnitude 4, which may be above the effective detection threshold of the earlier analogue seismic network. Thus, earthquakes of magnitude below M : 4, were either not consistently recorded or were not included in the catalogue, which is used to prepare the maps.

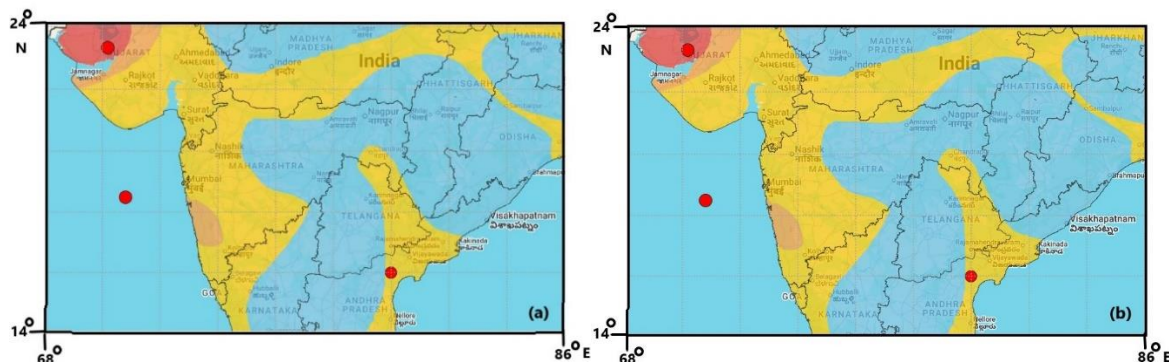


Figure 1. Location of earthquakes during 1950-1960. (a) M : 4-8, and (b) M : 1-8. Red solid circles indicate earthquakes with focal depth ≤ 30 km. Source: National Centre for Seismology Data Portal.

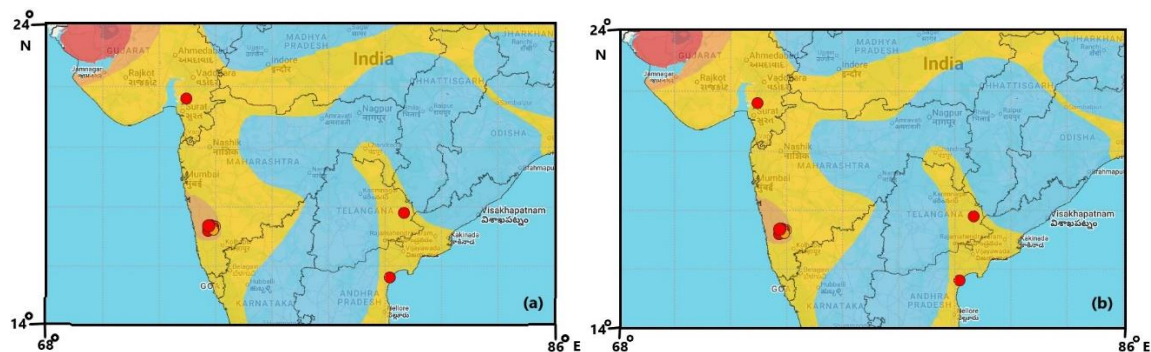


Figure 2. Location of earthquakes during 1960-1970. (a) M : 4-8, and (b) M : 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

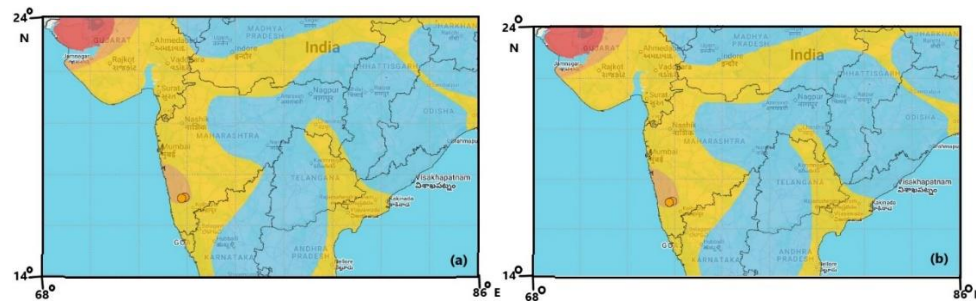


Figure 3. Location of earthquakes during 1970-1980. (a) M: 4-8, and (b) M: 1-8. Orange solid circles indicate earthquakes with focal depths 31-50 km. Source: National Centre for Seismology Data Portal.

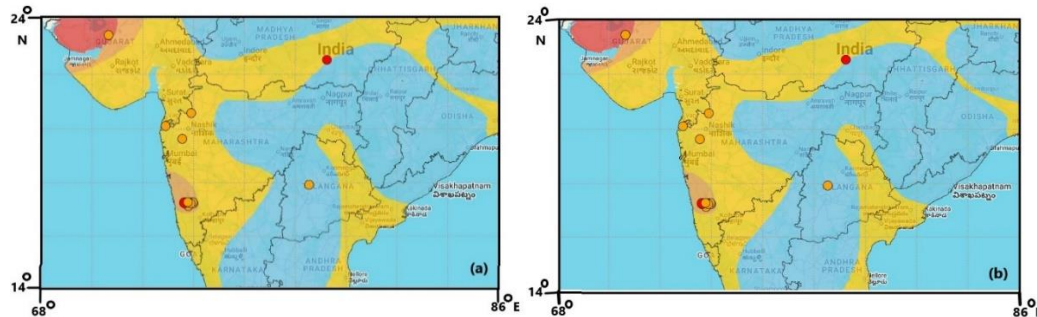


Figure 4. Location of earthquakes during 1980-1990. (a) M: 4-8, and (b) M: 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

Importantly, a distinct change becomes apparent after 1990 (Figures 5-8), coinciding with the progressive transition from conventional analogue recording to digital seismic acquisition and improved instrumental capabilities, that has improved substantially the detection ability of smaller earthquakes. Consequently, the magnitude 1-4 earthquake events become increasingly prominent after 1990, and the spatial distribution of seismicity becomes considerably more detailed. The maps for these periods reveal numerous small-magnitude events occurring around established seismic zones and along crustal weak zones. It is important to note that the devastating 1993 Killari earthquake fall in this interval.

The post-1990 period therefore shows two important characteristics. First, there is a clear increase in the total number of earthquakes in the M: 1-8 category, reflecting the growing number of detected events across the region. Second, there is a particularly pronounced increase in the M: 4-8 events, which becomes increasingly dominant in the earthquake catalogue. Most importantly, the continued occurrence of $M \geq 4$ earthquakes, together with their increasing spatial clustering in several parts of Maharashtra and adjoining regions, suggests that the temporal evolution cannot be attributed solely to instrumental improvements. The persistence of moderate earthquakes, provides an important basis for examining whether a genuine component of increasing tectonic activity accompanies the improvement in observational capability.

The temporal evolution becomes particularly significant during the 2000-2010 and 2010-2020 intervals (Figures 6 and 7), when

earthquake activity becomes progressively more widespread across Maharashtra and adjoining regions, which include the destructive 2001 Bhuj earthquake. Seismicity is observed not only around established zones such as Koyna, Killari and Bhuj, but also across the broader Marathwada-Telangana region and along several ancient structural trends. The increasing number of smaller earthquakes, provides a more complete representation of the active seismic fabric of the crust. The spatial distribution of these events suggests that the present-day seismicity is preferentially concentrated around crustal weak zones, including ancient faults, shear zones and rift-related structures.

The final 2020-2026 time window (Figure 8), shows a particularly notable concentration of seismic activity and provides an important contrast with the earlier decades. In addition to the continued abundance of low-magnitude earthquakes, the region experienced several significant moderate events, including the 2024 Mulugu earthquake and the recent Nanded-Parbhani-Hingoli seismic sequences in the Maharashtra region. The occurrence of these events within a relatively short time interval highlights the renewed seismic activity of the region and emphasizes the importance of examining the present-day behaviour of ancient crustal structures beneath the Deccan Volcanic Cover. Further, it provides a valuable opportunity to assess whether these events represent isolated episodes or part of a broader phase of intraplate deformation.

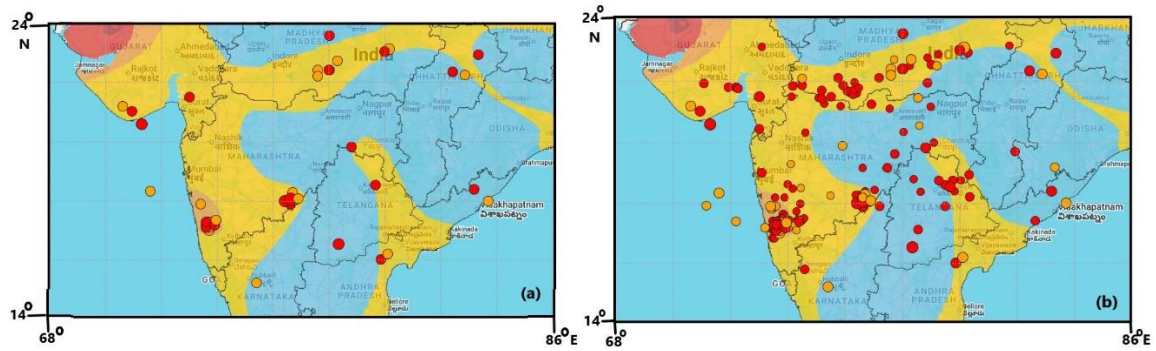


Figure 5. Location of earthquakes during 1990-2000. (a) M: 4-8, and (b) M: 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

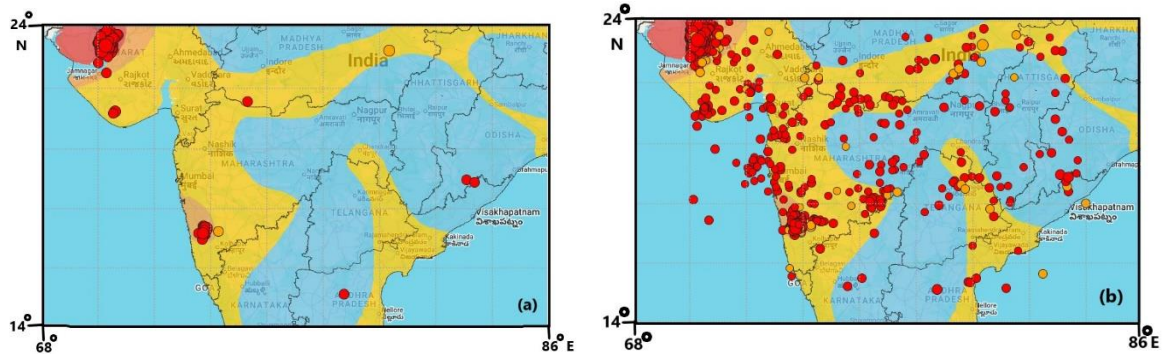


Figure 6. Location of earthquakes during 2000-2010. (a) M: 4-8, and (b) M: 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

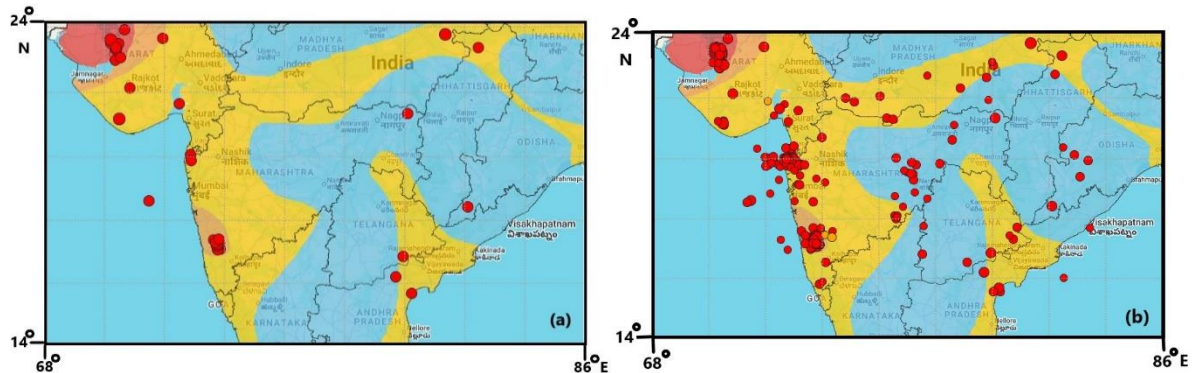


Figure 7. Location of earthquakes during 2010-2020. (a) M: 4-8, and (b) M: 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

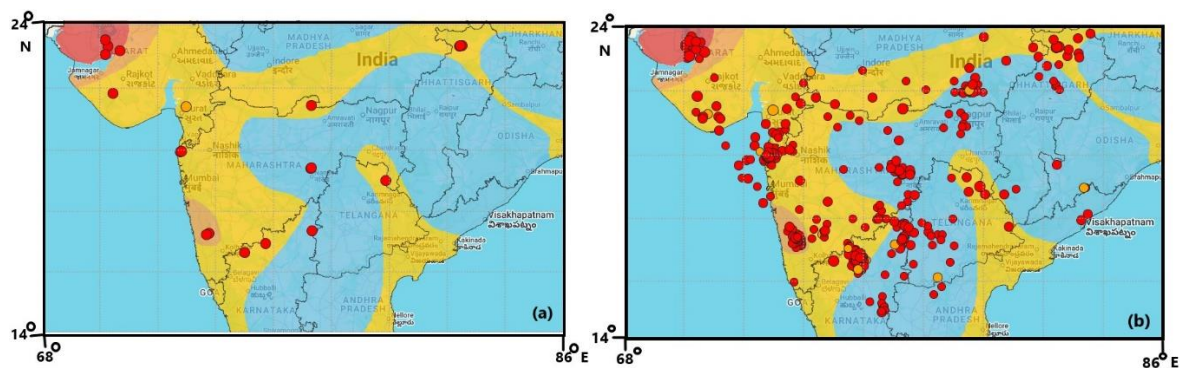


Figure 8. Location of earthquakes during 2020-2026. (a) M: 4-8, and (b) M: 1-8. Solid circles indicate earthquake locations; red circles represent focal depths ≤ 30 km and orange circles 31-50 km. Source: National Centre for Seismology Data Portal.

This temporal pattern provides an important foundation of evaluating the relationship between contemporary seismicity and the reactivation of inherited Precambrian faults and intra-cratonic rift systems, like Koyna, Kurduwadi, Godavari, Kutch and Cambay, under the influence of the present day regional stress field.

DISCUSSION

The recent seismic occurrences in the Deccan Volcanic Province, particularly in and around Nanded, Parbhani and Hingoli regions, became especially significant when considered together with the 2024 Mulugu earthquake and the long-term seismicity around Koyna, Killari and Kutch regions. The temporal analysis presented in this study, reveals a systematic increase in the seismic activity in the Maharashtra and adjoining regions. From this study, it is easy to understand the apparent clustering of recent earthquakes along inherited structural trends, which would suggest that the present-day Indian lithosphere is not mechanically homogenous. Instead, ancient faults, shear zones and rift related discontinuities, may continue to act as zones of weakness, capable of accommodating small increments of contemporary deformation.

Influence of far-field Himalayan compression and reactivation of ancient rift systems

The north-ward movement of the Indian Plate and its continued convergence with the Eurasian region, generates substantial compressional stresses in the Himalayan region. These stresses are not necessarily confined to the plate boundary, but can be transmitted southward into the Indian Plate in the form of broad-scale stress waves, or spatially varying stress fields. Regions underlain by strong and coherent lithosphere, may accommodate the stress elastically, whereas ancient rifts, shear zones and fractured basement domains, may preferentially accommodate deformation through fault reactivation. In this context, the Koyna, Kurduwadi, Godavari, Cambay and Kutch rift systems, can be regarded as potential stress-prone zones within the Indian Plate. The recent seismicity in Maharashtra and adjoining regions, may therefore represent localized release of accumulated stresses in such zones.

The occurrence of the 2024 Mulugu earthquake, along the broader Godavari structural domain (Vasanthi, 2025), provided a significant indication that this ancient rift system continues to influence contemporary deformation. The Nanded-Parbhani and Hingoli earthquake sequences, provide an additional important observation in this context. The proximity of the Hingoli main shock and its aftershocks to the NW-SE trending Kinnerasani-Godavari Fault, coinciding with western flank of Godavari graben, suggest that this seismic corridor may have

been reactivated or may have influenced the distribution of contemporary strain. The relationship between Mulugu, Nanded, Parbhani and Hingoli earthquakes, may therefore represents a broader pattern of episodic reactivation along the interconnected structural fabric of central India. Rather than considering these earthquake sequences as isolated events, their distribution suggest that stress may be transferred and redistributed between adjacent basement faults. Such a mechanism could produce spatially separated earthquake clusters at different times, depending on the geometry, orientation and frictional strength of individual fault segments. The observed seismicity may consequently represent a process of progressive adjustment of the heterogeneous Indian lithosphere to the contemporary stress regime.

Quaternary uplift, asthenospheric upwelling and lithospheric weakening

The Deccan volcanics and surrounding regions are characterized by significant variations in topography, crustal thickness and lithospheric structure. Almost the entire peninsular shield has been associated with continuous uplifting and destructive erosion since the early Tertiary until the present. The area includes the Deccan Trap region, which also contains several prominent Quaternary uplifted regions, like Sangola, Manmad-Sangamner, Aland etc. (Radhakrishna, 1993). The presence of these major uplifted domains in the Deccan region, may indicate spatial variations in lithospheric buoyancy beneath the underlying asthenosphere, and this continuous uplifting reactivates pre-existing faults that cause tensile fracturing on either side of the uplifting blocks. The movement of blocks along these reactivated faults may be partly responsible for the continuous seismicity of the Deccan volcanic region, which contains one of the most faulted/fractured and regionally displaced upper crustal blocks (Ravi Shanker, 1995).

Such uplift cannot automatically be attributed to the present-day mantle upwelling, but its spatial correspondence with geophysical anomalies, provides a basis for considering a possible thermal component in the regional geodynamics. Geophysical observations from the broader Deccan-Kutch-Cambay regions, have suggested anomalous upper-mantle velocities, thermal perturbations and evidence for partial melting or magmatic underplating at shallow depths. In certain zones, like the Godavari and Son-Narmada rifts, parts of the Cambay graben, and the western faulted margin of the Konkan coast, temperatures at a depth of 500 m, exceed $65\pm 10^\circ\text{C}$ (Agrawal and Pandey, 1999). These are the areas where enhanced seismic activities are being experienced now. Also, Quaternary uplifting and large lateral and vertical crustal movements are quite common here, as the asthenosphere beneath the Indian shield has been thermally remobilised by

plume interactions several times in the past. Thus, the unusual tectonic activity within the Indian shield, appears to stem from the hot conditions underneath, where local stresses are continuously building up due to a rise of isotherms caused by hot and upwarped asthenosphere (Agrawal and Pandey, 1999).

The asthenospheric upwelling persisting beneath the study region, could probably produce dynamic support and spatial variations in lithospheric temperature and strength. Thermally weakened portions of the lithosphere, would be expected to respond differently to regional compression than mechanically stronger domains. The proposed role of asthenospheric upwelling is not that it directly generates individual earthquakes, but it possibly modifies the mechanical strength and stress distribution of the lithosphere, thereby influencing the faults, which are most susceptible to reactivation. In this sense, mantle thermal processes may act as a background control that enhances the response of pre-existing faults to far-field tectonic stresses. The future direction of this present study could be integration of seismicity, GNSS, gravity, heat flow and other geophysical data. Such an integrated dataset would allow the relative contributions of far-field tectonic stress, inherited crustal architecture and deep thermal processes to be quantitatively evaluated.

CONCLUSIONS

The present study highlights a marked increase in seismicity across Deccan volcanic and adjoining regions over the past several decades. This increased seismic activity, can be interpreted as combined influence of far-field compressional stresses associated with the India-Eurasia collision, redistribution of stresses within the Indian lithosphere, and progressive reactivation of inherited crustal weaknesses associated with Koyna, Kurduwadi, Godavari, Kutch and Cambay rift systems. In addition, Quaternary uplift and possible asthenospheric thermal upwelling beneath the Deccan volcanic and adjoining regions, may contribute to lithospheric weakening and modification of the regional stress field, thereby facilitating fault reactivation.

Acknowledgements

I sincerely thank Dr. Prakash Kumar, Director, CSIR-National Geophysical Research Institute, Hyderabad, for the permission to publish this work. I profoundly thank the Editor and the Reviewers for their valuable comments and suggestions to improve the quality of the manuscripts. This is CSIR-NGRI Contribution No - NGRI/LIB/2026/Pub-130.

Author Credit Statement

The total work is carried out solely by the corresponding and submitting author in this paper.

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The author does not have any competing interest and adhere to copyright norms.

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Complex disaster dynamics in the Himalayas: a case study of the August 2025 Kheer Ganga flood event

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ABSTRACT

The Indian Himalayan Region faces frequent natural disasters from heavy rainfall, cloudbursts, glacial lake outburst floods (GLOFs), glacier retreat, and landslides. Mountain communities are highly vulnerable due to hydroclimatic extremes, steep slopes, unstable debris, and fast-flowing rivers. This study examines the August 2025 flood in the upper Bhagirathi valley, Uttarkashi district of the state of Uttarakhand, India, focusing on the Dharali–Harsil area near, the Kheer Ganga River. Initial reports linked the event to a cloudburst; however, rainfall intensities from 3–5 August (25.44–45.64 mm/hr) were less than 100 mm/hr cloudburst threshold. The data suggests prolonged and uneven rainfall, combined with other triggers such as GLOFs, glacier collapse, or landslide. This case underscores the need for integrating meteorological, hydrological, and remote sensing data to improve hazard mapping at micro-level, early warning systems, and climate-resilient planning for vulnerable Himalayan communities.

Keywords: Flash flood; Cloudburst; GLOFs; Glacial Lake; Debris flow; Himalayan hazards; Kheer Ganga River; Uttarakhand

INTRODUCTION

Natural disasters such as extreme rainfall and cloudbursts, pose significant threats to both human life and the environment, with impacts that can persist for many years (Sain, 2026a, b). These challenges are particularly severe in mountainous regions, where steep slopes, unstable terrain, and swift river flows, magnify the destructive potential of such events. In these areas, disasters can damage infrastructure, destroy agricultural land and forests, disrupt livelihoods, and hinder long-term community development. The Indian Himalayan Region is a critical resource base for the country, providing water, hydropower, and agricultural support. However, its complex topography and fragile ecosystems, make it highly vulnerable to hazards such as floods, cloudbursts, glacial lake outburst floods (GLOFs), and landslides (Dash and Punia, 2019; Chand et al., 2022). Most of the cloudbursts in the Himalayas, occur during the peak monsoon season (Bhan et al., 2004; Dimri et al., 2017; Yadav et al., 2020; Mishra et al., 2022). Deshpande et al. (2018) indicated that cloudbursts are not random events but are embedded within mesoscale cloud systems, often taking place in the early morning hours along the Himalayan foothills.

The past decade has witnessed several high-impact events in Uttarakhand, including the 2013 and 2021 disasters, where intense rainfall triggered widespread flooding and landslides (Nandargi et al., 2016; Rawat et al., 2022). The June 2013 Kedarnath tragedy, one of the worst disasters in the region since the 2004 Indian Ocean tsunami, was caused by a combination of a cloudburst, persistent heavy rainfall, and a GLOF, resulting in severe flash floods and landslides (Dobhal et al., 2013; Bhambri et al., 2016; Patel et al., 2022). Meteorological analysis by Pratap et al. (2020) revealed that high humidity, dense cloud cover, low temperatures, and strong winds during

this period promoted rapid cloud condensation and precipitation. Recently, Nagamani et al. (2024) investigated the rare October 2021 flash floods in Nainital, Uttarakhand, using remote sensing and statistical techniques. The study highlighted the influence of both atmospheric conditions and local geographic factors, emphasizing the growing role of climate change in intensifying such events. Remote sensing, morphometric parameters, and Geographic Information System (GIS) tools, have been increasingly employed to create flash flood hazard maps, aiding in the identification of vulnerable zones and supporting disaster risk reduction strategies (Sahu and Rawat, 2024). This study focuses on the flood event that occurred in Uttarkashi from 3–5 August, 2025. The research investigates the potential causes of this event, evaluates its underlying meteorological and geomorphological factors, and discusses its implications for disaster preparedness and mitigation in the IHR.

STUDY AREA

This study was carried out in the upper Bhagirathi valley in Uttarkashi district (Figure 1), Uttarakhand, India. The study area is mainly situated around Dharali village and the nearby Harsil area. This is a steep mountaineous region shaped by glaciers, with narrow valleys that are either V-shaped or U-shaped. Uttarkashi district covers an area of 8016 sq.km. with elevation ranging from 1000 m (lower valleys) to 7800 m (high Himalayan peak). Slope of the area also varies between gentle slopes (<15°) and very steep slope (>45°) in valley bottoms and upper mountainous regions respectively. Mean annual temperature of the region ranges between 5 to 20°C with 1000 to 1500 mm avg. annual rainfall. At the meeting points of smaller side streams, there are large deposits of rocks and soil from past glacier and river activity. Dharali and Harsil are on

the right side of the Bhagirathi River, close to where smaller but fast-flowing streams like Kheer Gad (also called Kheer Ganga) and Jalandhari Gad join the river. Many houses and roads here are built on these rocky deposits.

DATASETS AND METHODOLOGY

Rainfall data were collected from the NASA Giovanni platform using the Global Precipitation Measurement (GPM) product. The GPM IMERG product has a nominal spatial resolution of $0.1^\circ \times 0.1^\circ$ (approximately 10–11 km in latitude, with a smaller east–west dimension at Himalayan latitudes). The Integrated Multi-satellitE Retrievals for GPM (IMERG) algorithm combines observations from multiple satellites within the GPM constellation to estimate precipitation. For this study, half-hourly rainfall data were used. The precipitation data for the

flood period were downloaded in NetCDF format and processed for the study area.

The August 2025 Kheer Ganga disaster

In the early August of 2025, a devastating natural disaster struck parts of the Himalayas, focusing heavily on areas along the Kheer Ganga River. The exact cause of the event is still under investigation by scientists, government agencies, and disaster management teams. Although there is agreement that the event involved a sudden and destructive surge of water, mud and debris, there is not yet a clear consensus regarding the trigger. The complexity arises because in mountainous regions, especially in the Himalayas, many different types of events both natural and climate-driven, can happen almost at the same time, and each can set off or worsen the others.

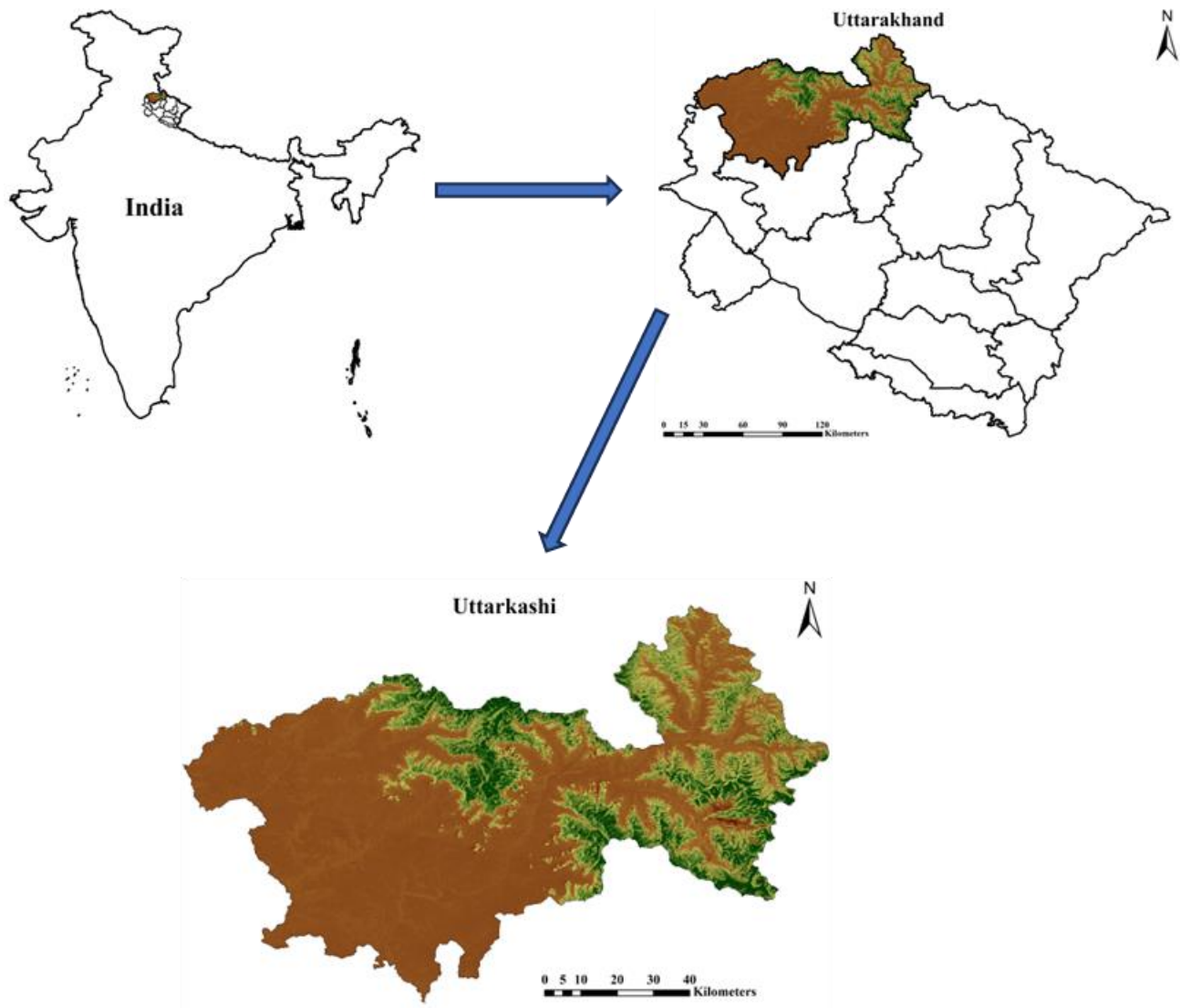


Figure 1. Location map of the study area in the upper Bhagirathi valley, Uttarkashi (Uttarakhand), India

Rainfall pattern

The rainfall data for Uttarkashi district in early August 2025, gives an important clue about the possible cause of the flood. The highest and lowest rainfall intensities were recorded each day from 3 August to 5 August 2025 (Figure 2). First, it is important to understand “what is a cloudburst?”. A commonly used operational criterion for a cloudburst in the Indian Himalayan context is rainfall exceeding approximately 100 mm h^{-1} over a localized area within a short period. The GPM-IMERG data used here did not resolve rainfall rates above this value; however, the spatial resolution of the product limits the ability to exclude localized sub-grid extreme rainfall. Such events can release huge amounts of water suddenly, often triggering flash floods, landslides, and debris flows. The highest rainfall rate recorded during this period was 45.64 mm/hr on 5 August. The maximum rainfall rate resolved by the GPM-IMERG product was 45.64 mm h^{-1} . This value is below the commonly used 100 mm h^{-1} criterion for cloudburst classification. However, given the approximately 0.1° spatial resolution of the satellite product, these data cannot by themselves exclude a highly localized cloudburst near the Kheer Ganga catchment. Ground-based observations and higher-resolution precipitation datasets are required for definitive attribution. The other days, 3 August and 4 August, also had high rainfall rates of 25.44 mm/hr and 39.43 mm/hr ,

respectively again, strong but not extreme enough to qualify as a cloudburst. The lowest rainfall readings on each day, between 0 mm/hr and 2.8 mm/hr , show that not all parts of the district experienced the heavy rainfall at the same time. This variation suggests the rain was unevenly distributed across different locations. Because cloudbursts are highly localized, their detection requires sufficiently dense spatial and temporal observations. The spatially coarse satellite data used here may smooth localized precipitation maxima, overwhelming rivers and drainage channels almost instantly.

This pattern of moderate-to-heavy, but not extreme rainfall over several days, points toward a different kind of flooding process. Instead of a sudden single burst of extreme rain, the data suggests prolonged and uneven rainfall. This can cause rivers to gradually swell, especially when combined with other factors such as, rapid snowmelt or glacier ice melt caused by warm temperatures, the sudden release of water from a glacial lake (a GLOF event), and landslides or debris dams in river valleys that temporarily block water and then collapse, sending a surge downstream. Such processes can occur alongside heavy rainfall but are not triggered solely by it. In the case of the August 2025 Uttarkashi flood, the rainfall intensity data makes it unlikely that a cloudburst was the main cause. Instead, the flood may have been driven by a combination of moderate-to-heavy rain and one or more other mountain hazards.

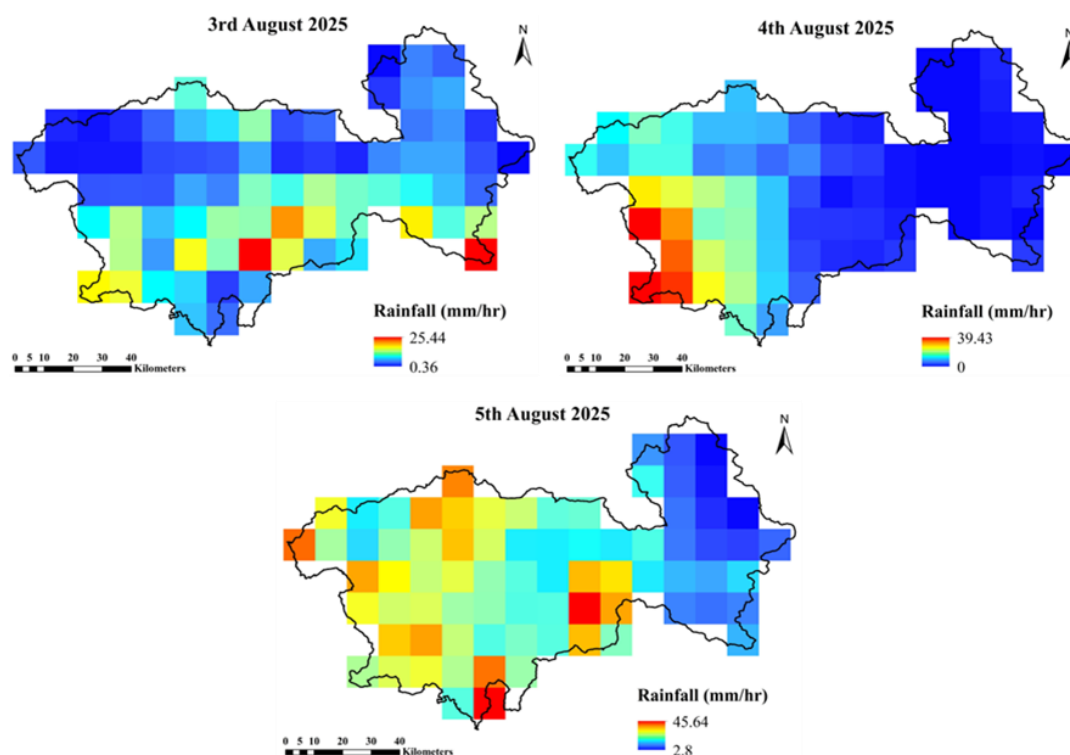


Figure 2. Rainfall pattern over the study area during 3rd to 5th August, 2025

Initial reports and cloudburst theory

When the disaster first happened, media outlets quickly reported that a cloudburst might have been the main cause. As stated earlier, a cloudburst is an extreme and sudden downpour, where a massive amount of rain falls in a very short time often within minutes or hours over a small geographical area. In mountainous regions, this kind of rainfall can be devastating because steep slopes and narrow valleys cannot absorb so much water quickly. The rainwater rushes downhill at high speed, picking up soil, rocks, and debris, and often turning into a flash flood. From a journalist's perspective, a cloudburst was a logical early guess. It is a well-known phenomenon in the Himalayan states, particularly during the monsoon season. In many past cases such as in Leh (2010) or Kedarnath (2013), cloudbursts have been linked to massive destruction. Moreover, the India Meteorological Department (IMD) had already predicted heavy rainfall in the region until 10 August, which seemed to support this theory. However, early media reports are often based on quick observations, eyewitness accounts, and local information. These can be helpful for initial awareness, but may not capture the full complexity of what actually happened.

Re-examining the cause: Other possible triggers

As scientists began to study the event more closely, they realized that a cloudburst might not be the only explanation or even the main one. Instead, other geological and hydrological possibilities came into focus.

Glacial Lake Outburst Flood (GLOF)

In high-altitude areas, glaciers often form small lakes at their edges or within depressions created by melting ice. These are called glacial lakes. They are usually held back by walls of ice, rock, or loose sediment. Over time, due to warming temperatures, these lakes can grow larger and more unstable. If the natural dam holding the water breaks because of melting ice, a landslide, or an earthquake large volume of water can rush downstream. This is called a GLOF. GLOFs are particularly dangerous because they happen suddenly and carry massive energy, sweeping away almost everything in their path. Given the Himalayan region's vulnerability to glacier melt (exacerbated by climate change), it is now being explored whether a glacial lake above the Kheer Ganga area might have burst, releasing a wave of water that caused the flooding.

Glacier Collapse

Sometimes, instead of the lake bursting, an actual chunk of a glacier breaks away from the main body and crashes downhill.

This can happen when a glacier becomes unstable due to warming or due to heavy rainfall weakening the ice structure. A collapsing glacier can directly release meltwater or trigger landslides and dam breaks, leading to similar destructive flooding. The 2021 Chamoli disaster in Uttarakhand is an example where a glacier fragment collapse set off a chain reaction of flooding and destruction.

Landslide-Induced Flood

Landslides are common in the Himalayas due to steep slopes, unstable soil, and heavy rainfall. A landslide can block a river temporarily, forming a natural dam. When the dam breaks whether minutes or hours later, water rushes downstream with extreme force. Landslides can also dump massive amounts of debris into rivers, raising water levels and creating flooding upstream and downstream.

Combination of events

In reality, disasters in mountain areas often result from a chain of interconnected events (Figure 3). For example, heavy rain could trigger a small landslide, which destabilizes a glacier, causing ice to fall into a glacial lake, which then bursts and floods the river. Researchers are keeping this "multi-trigger" possibility open, as multiple factors may have worked together to cause the Kheer Ganga flooding.

THE DISASTER'S IMPACT

What is clear is that whatever the exact cause, the Kheer Ganga River overflowed with immense force. The water was not just clear river water, it carried large amounts of debris, mud, and possibly rocks and vegetation ripped from the slopes. Such a mixture turns a normal flood into a much more destructive event known as a debris flow or mudflow. In details, debris flow refers to a rapid downslope movement of a water-saturated mixture of soil, rock fragments, and organic material, while mudflow refers to a fine-grained, sediment-rich flow dominated by silt and clay-sized particles. These are heavier and more forceful than clean water floods because the debris increases the density and destructive power of the flow.

Floodwaters inundated and severely damaged Dharali village and nearby areas, damaging homes, sweeping away roads and bridges, and affecting farmland and local infrastructure. For the people in the path of the flood, the event would have been terrifying, giving little time to react. Heavy debris in the water can also make rescue and relief work much harder, as it blocks access roads and makes rivers dangerous to cross.

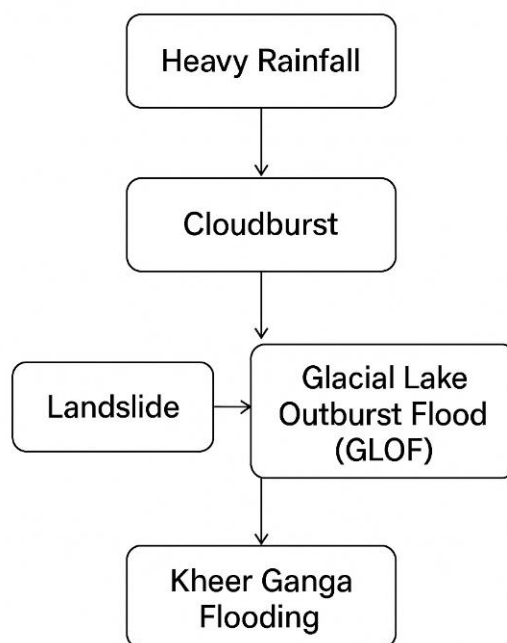


Figure 3. Possible sequence of events leading to Kheer Ganga flooding

GEOLOGICAL PERSPECTIVES

Himalaya - a most fragile and youngest mountain system in the world, is characterized by complex geological and tectonic settings. Recently a devastating flash flood occurred due to cloudburst and torrential rainfall in the narrow gorges of Kheer Ganga (a tributary of Bhagirathi River) in Dharali and Harsil villages of Uttarkashi district in the higher Garhwal Himalaya. Geologically this terrain is traversed by low, middle and high-grade metamorphic rocks (crystallines). These crystalline rocks are tectonically separated by Main Central Thrust (MCT), Jutogh and Vaikrita thrusts (Valdiya, 1980). Later these tectonic boundaries are designated as MCT₁, MCT₂ and MCT₃ of a complex duplex system by Bahuguna and Saklani (1988). This complex duplex system, reveals a relationship between the minor duplexes with the major Central Crystalline Duplex and is investigated with the help of branch-line maps and balanced cross-sections (Bahuguna and Saklani, 1988). These low to high grade metamorphic units exhibit steep dips and represented by highly sheared and mylonitised rocks (Bhatt and Saklani 1990, 1994). Dharali belongs to higher Himalaya and constitute medium to high grade crystalline rocks.

The Kheer gad/stream, a small tributary of Bhagirathi River, dumped large amount of debris and alluvial sand and reshaped its course and formed alluvial fan (Figure 4) at the confluence of Bhagirathi River during catastrophic flash flood that occurred on 5 August 2025. The stream channel has been widened, buried several buildings and human lives and formed

alluvial fan in approximately ~20 ha area. The encroached houses settlements built on flood plains and alluvial fan, were washed away and many people died and there was huge loss of property and livestock (Figure 4a, b). The flood has obstructed at the meandering point (Figure 4a) and has increased the intensity of the destruction of houses built close to the curved meander during this disaster.

The illegal construction and encroachment of flood plains have become common phenomena in the Himalayan region. Nobody is following the law of river morphology (i. e. settlement should be at least 200 meters away from the river channel) and deliberately encroaching the river channel and flood plain. This approach is being followed by people in the Mandakini valley and other regions also even after the catastrophic Kedarnath tragedy. A catastrophic flash flood occurred in the territory of Kedarnath shrine and Rambara, Gaurikund, Sonprayag, Kalimath and other towns located in the downstream of Mandakini River on 15 and 16 June 2013 and caused huge damage to human life and property. The central and Uttarakhand state governments should issue directives to follow the environmental laws. The extreme weather events (cloudbursts, flash floods, landslides and earthquakes etc.) were occurring in the past and likely to occur in the future in these seismically sensitive and vulnerable zones of cloudburst, flash flood and landslides. The intensity may be higher due to climate change and due to other reasons but the life and property can be saved by avoiding unplanned and illegal construction/urbanization/settlements.



Figure 4. Satellite images of Dharali village showing picture of (a) before, (b) after flash flood, released by Indian Space Research Organization (ISRO). Dharali settlement occupies portions of an alluvial-fan/floodplain surface and (c) The Dharali town which was completely washed away after flash flood and alluvial fan of Khir Gad. This photo shows Kheer Ganga where formation of alluvial fan took place after dumping the huge amount of debris load that reshaped its old channel/alluvial fan during flash flood

DISCUSSION

Before the disaster, the IMD had issued a warning for heavy rainfall in the region, valid until 10 August. Such warnings can help local authorities prepare for possible flooding, landslides, or other weather-related hazards. They can alert communities to stay indoors, avoid riverbanks, and prepare emergency supplies. Heavy rainfall is not only a hazard by itself (through flooding and cloudbursts) but can also act as a trigger for GLOFs, glacier collapses, or landslides by saturating the soil, melting ice, and adding weight and pressure to unstable slopes.

While the warning did not prevent the disaster, it provides a clue that extreme weather was part of the sequence of events that led to the flooding.

Why the further investigation matters

Finding out the exact cause of the disaster is important not only for science but also for protecting lives and property in the future. If the disaster was caused by a GLOF, authorities will need to watch glacial lakes more carefully and set up early warning systems to prevent sudden floods. If it was mainly due

to a landslide or heavy rainfall, then better monitoring of slopes and improved water drainage systems can help reduce the risk. Understanding the cause can also help educate local communities about the specific dangers in their area, so they can respond quickly and evacuate in time. It will guide the government in making decisions about where to allow construction, where to strengthen riverbanks, and where to limit human activities. In the fragile mountain regions of the Himalayas, where tourism, hydroelectric projects, and settlements are growing, such knowledge is essential for planning safer and more sustainable development.

CONCLUSIONS

The Kheer Ganga disaster is a reminder of how complex and interconnected natural hazards can be, especially in high-mountain regions. The available evidence suggests that the event may involve multiple interacting hydrometeorological and geomorphological processes; however, the present analysis cannot establish the relative contribution of individual triggers. Regardless of the final verdict, the flood's destructive power was fuelled by large amounts of debris and mud, making it especially dangerous for communities in its path. The IMD's earlier heavy rainfall warning shows that extreme weather was part of the equation, but the exact trigger may involve deeper geological and climate-related processes. For the people of Dharali and surrounding areas, the priority now is recovery and rebuilding. For scientists and disaster managers, the task is to learn from this events that early warnings, monitoring systems, and preparedness measures can be strengthened, potentially saving lives in the future.

Author Credit Statement

Kishan Singh Rawat: Conceptualization, methodology, supervision, project administration, writing – review and editing. Smruti Ranjan Sahu: Data curation, formal analysis, investigation, visualization, methodology, writing original draft. Sudhir Kumar Singh: Validation, resources, data interpretation, writing, review and editing. S.C. Bhatt: Supervision, validation, funding acquisition, writing, review and editing. All authors have read and approved the final manuscript.

Data availability

Data is available on reasonable request from corresponding author

Compliance with ethical standards

There is no conflict of interest and authors adhere to copyright norms

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Received on: 13-12-2025; Revised on: 25-06-2026; Accepted on: 25-06-2026

ERRATUM

Authors erratum to the paper “**Shallow upper crustal structure of south Rewa Gondwana basin, Central India constrained from different geophysical studies**”, J. Ind. Geophys. Union, 30(4) (2026), 241-252
([https://doi.org/10.71122/JIGU.30\(4\)2026.0024](https://doi.org/10.71122/JIGU.30(4)2026.0024)).

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The conclusions of the paper are unaffected by this correction.

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Accepted on 14-08-2026

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